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JOURNAL OF
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RESULTS OF HADDOCK TAGGING IN THE GULF OF MAINE FROM 1923 TO 1932[1]

By

WILLIAM C. SCHROEDER

In the hope of throwing some light on their migratory habits 10998 haddock were tagged in the Gulf of Maine from 1923 to 1932 by the United States Bureau of Fisheries vessels *Halcyon* and *Albatross II*, during the course of cod marking experiments, made between the months of April and October. This paper gives an account of the recaptures reported from these tagged haddock, together with a brief discussion of rate of growth.

What little was known of the haddock's migrations in the Gulf of Maine up to 1924 is summed up by Bigelow and Welsh (1925) as follows: "Adult haddock roam from place to place in search of food like cod, and so constantly that where there is good fishing today, there may be none tomorrow. However, their movements seem mostly of short extent, from place to place on a given bank as food is locally exhausted or from some other cause. How much interchange of haddock there may be from bank to bank or between inshore and offshore grounds is unknown, but the fish that inhabit the coastwise belt carry out a local and irregular migration inshore in winter and early spring and offshore again in June or July. Certain bodies of fish may linger all summer in the deeper channels among the islands of Maine, on patches of suitable bottom."

[1] Published with the approval of the Director of the Fish and Wildlife Service, U. S. Dept. of the Interior.

Contribution No. 309 of the Woods Hole Oceanographic Institution.

Needler (1930) discusses the migrations of haddock in the northwestern Atlantic basing his conclusions on about 7000 fish tagged by the Biological Board of Canada and about 8000 fish which, up to that time, had been tagged by the U. S. Bureau of Fisheries. It is these latter 8000 haddock, together with some tagged later, that form the basis of the present report. Any discrepancies, as to numbers of fish tagged and recaptured between Needler's report and this one are due to certain changes in the records which I originally furnished to Dr. Needler and to additional data obtained since then. Perhaps the only important error in the figures is that of the 6000 haddock tagged by the Bureau of Fisheries in the vicinity of Mt. Desert, as stated by Needler (1930, p. 28), but which should have been about 4000.

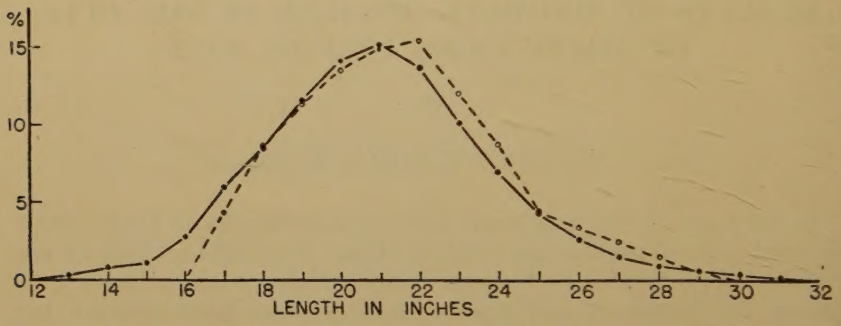


Figure 1. Length frequency distribution by percentages. Solid line: lengths of 10,998 haddock at time of tagging. Broken line: lengths of 210 recaptured haddock, at time of tagging.

Needler concludes from these haddock tagging experiments that the haddock population in the shallow waters along the New England coast does not mix with that of the Nova Scotia region. And none of the Nova Scotian tagged haddock were reported recaptured in the coastal waters or offshore banks of the New England coast hence there is no evidence of interchange between these regions.

The methods of tagging these haddock were identical to those used for cod by Schroeder (1930). In most cases metal tags were attached to the base of the tail dorsally. The tail of the haddock being rather tender many tags dropped off within the first 3 to 6 months, perhaps as many as 70 to 80 per cent. It is probably largely because of this that only 210 fish, or 1.91 per cent of the 10998 that were tagged were reported recaptured, a relatively small return as compared with about 12 per cent for cod marked on the same grounds.

Most of the haddock were tagged in the Nantucket Shoals region and in the vicinity of Mt. Desert, Maine. Smaller numbers were in-

cluded in the experiments in the shore waters between Cape Cod, Massachusetts, and Boothbay Harbor, Maine, and on the offshore banks (Platts Bank, Cashes Ledge, Browns Bank, and Georges Bank).

All of the fish were caught with hand lines and in size most of them fell between the lengths of 17 and 25 inches. Only 4.9 per cent of the catch was below 17 inches in length. No recaptures were reported from the smaller fish but beginning with those of 17 inches the size distribution of the recaptured fish agrees closely with the size distribution of all tagged fish (Figure 1).

NANTUCKET SHOALS REGION

TABLE I—Record of Haddock Tagged in the Nantucket Shoals Region, with Subsequent Recaptures

Year	Number tagged	Local Recaptures	Distant Recaptures	No data	Total Recaptures	Percentage
1923	410	4		2	6	1.46
1924	778	6	3	1	10	1.29
1925	1821	14	4		18	1.00
1927	481	2			2	0.42
1928	239	5	1	2	8	3.35
1929	132	3	1		4	3.03
1930	85		1		1	1.17
Total	3946	34	10	5	49	1.24

Nearly all these fish were tagged between Orion Shoal and Great Rip buoys; a few were tagged about 10 miles northeast of Orion Shoal on the "Chatham Grounds," and about 20 miles northeast, in South Channel. Recaptures of Nantucket Shoals region fish reported from any of these grounds are here classed as local. Precise recapture localities were seldom given by the fishermen, so it was not possible to trace with certainty a migration from Nantucket Shoals to nearby South Channel or vice versa.

Of the 34 local or nearby recaptures, 10 were taken in 20–100 days, 8 in 101–200 days, 6 in 201–300 days, 6 in 301–400 days, 3 in 401–500 days, and 1 in 601–700 days after tagging. The tagging vessels recaptured 3 fish 321, 330 and 377 days later, respectively. Thus, if it can be accepted that these few recaptured tagged haddock are at all representative of the bodies of fish of which they were a part, the indications are that a large proportion of the haddock population present at any given time in the Nantucket region remains localized for upwards of a year or more.

The 10 recaptures taken in distant localities prove that some of the haddock living in the Nantucket region carry out long journeys. Based on the total number of recaptures (49) and allowing that three of the "no data" fish were taken locally and two elsewhere, about one fourth of the Nantucket haddock migrated away. This is no doubt

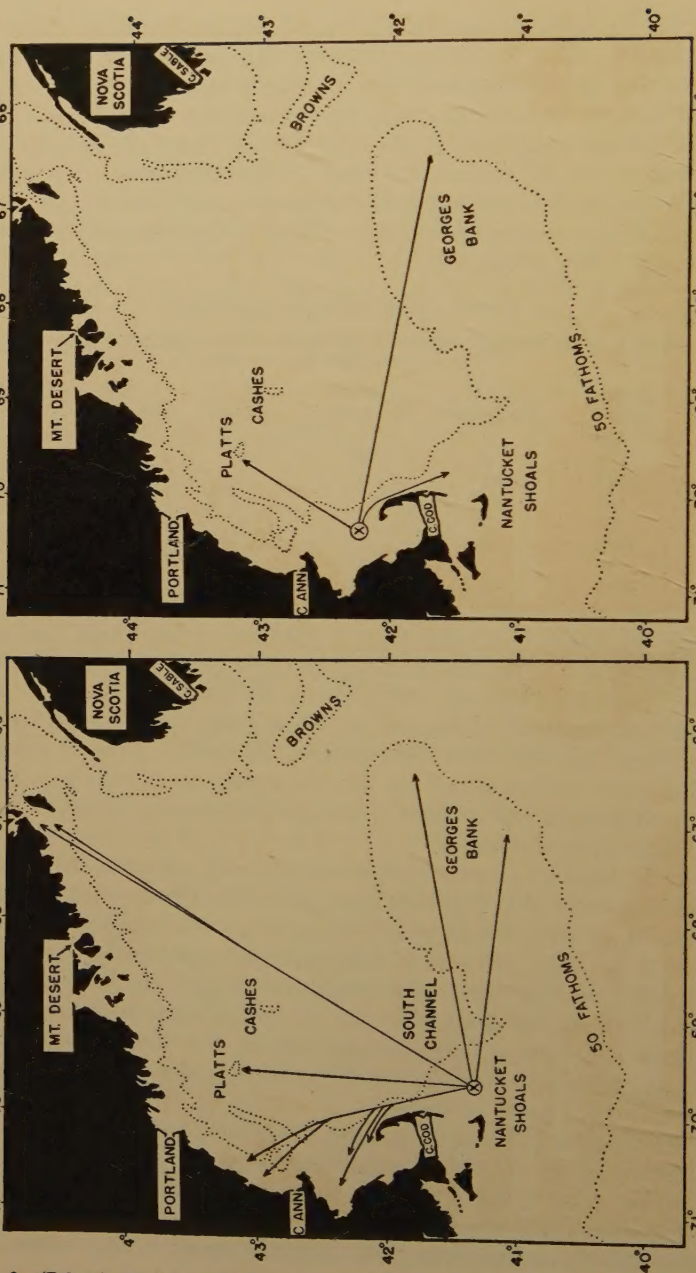
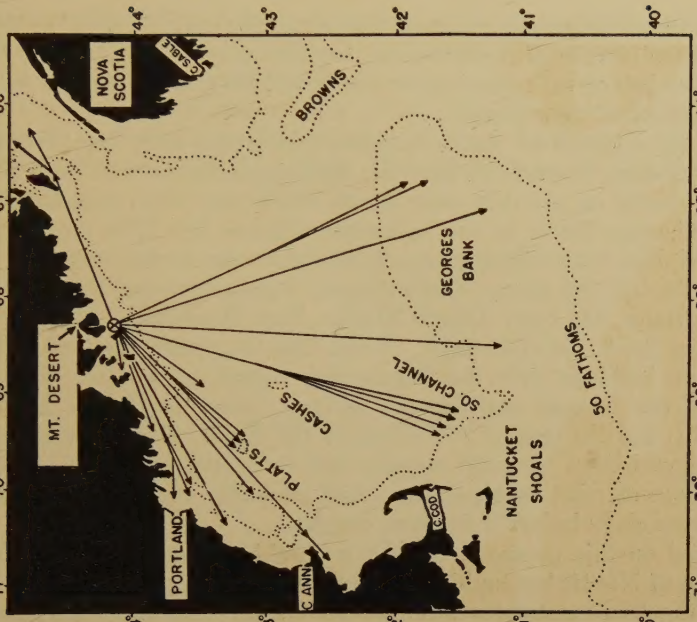


Figure 2. (Below.) Recaptures in distant localities of haddock tagged in the Nantucket Shoals region from 1923 to 1932.

Figure 3. (Above.) Recaptures of haddock tagged on Stellwagen Bank, from 1924 to 1928.



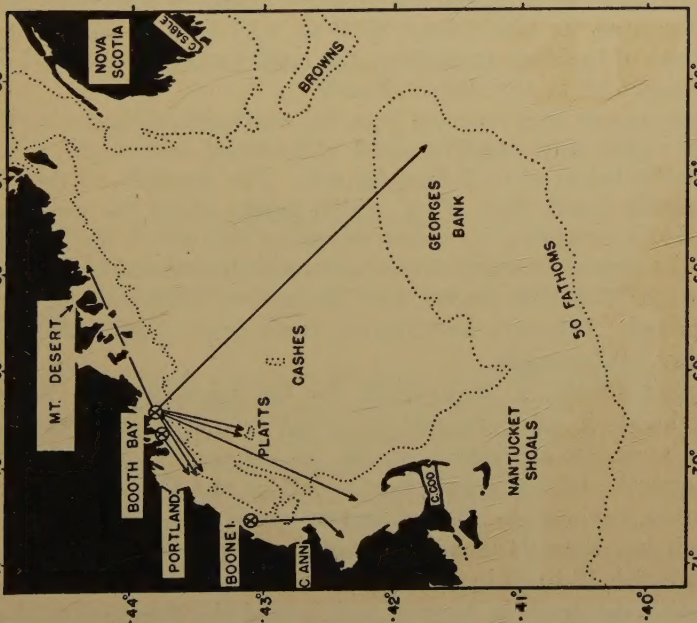


Figure 4. (Below.) Recaptures in distant localities of haddock tagged around Boone Island and Boothbay Harbor, Maine, from 1924 to 1930.

Figure 5. (Above.) Recaptures in distant localities of haddock tagged off Mt. Desert, Maine, from 1924 to 1932.

a minimum figure for, whereas the non-migrant fish stood a reasonably good chance of recapture by remaining in a region where fishing is rather intensive, some migrant haddock very likely moved to localities seldom if ever fished.

Of the migrant or wandering haddock eight were reported to the northward of Cape Cod, between Stellwagen Bank and New Brunswick, chiefly on grounds near shore, while only two recaptures came from Georges Bank to the eastward (Figure 2). The localities and elapsed days between tagging and recapture are as follows: Stellwagen Bank 69, 94, 286 days; Newburyport, Massachusetts, 158, 424 days; Platts Bank, 131 days; Grand Manan, New Brunswick, 105, 292 days; Georges Bank 114, 136 days. From this it appears that most of the migrant haddock left the Nantucket region soon after they were tagged the longest and fastest journey being to Grand Manan, a straight line distance of 240 miles, in 105 days. For the migrant fish the elapsed days between tagging and recapture can be considered a maximum as most or all of them had probably reached their destination some days before they were caught.

Based on this meager evidence it would appear that a larger proportion of Nantucket haddock go northward than go eastward. However, the two recaptures reported on Georges Bank to the eastward are significant for, although an intensive fishery is conducted on parts of this ground, its area (about 8500 sq. mi.) is much larger than the total area of haddock grounds to the northward of Nantucket Shoals from Cape Cod to the Bay of Fundy.

No recaptures were reported from the westward of Nantucket Shoals. In this region only small numbers of haddock are caught, and these are usually taken while fishing for cod. For example, in the vicinity of Montauk Point, New York, which marks about the western limit for haddock excepting for a few stragglers that reach the New Jersey coast, the aggregate catch of haddock made by several high line boats during the 1925–1926 season was: December, 4000 pounds; January, 2000 pounds; February and March, "very few"; April, 1000 pounds; and May, 8000 pounds.

Another source of information on the migrations of Nantucket Shoals haddock was furnished by length frequency distributions and by the apparent abundance of fish on successive trips made to the same grounds (Orion Shoal buoy and Rose and Crown buoy) during the spring, summer and fall of certain years. For example, in 1923, haddock less than 23 inches in length were virtually absent in May but some fish 17 to 22 inches long were taken in June, they comprised about half the catch in August, and formed the dominant size groups

in October (Figure 6). Thus there is good evidence that during the summer these 17–22 inch haddock moved to Nantucket Shoals, at least to that locality where tagging was done. From whence they came is unknown but the nearby South Channel grounds and the western part of Georges Bank, with a large haddock population, would seem to be a likely source. Whether many of the haddock

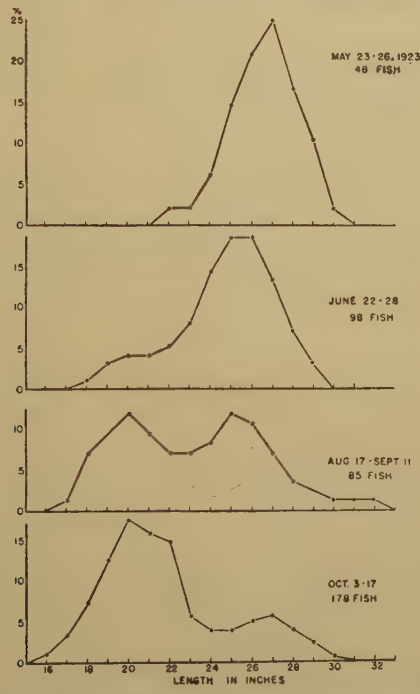


Figure 6. Length frequency distributions by percentages of haddock caught by the *Halcyon* on Nantucket Shoals during 1923. All length distributions are smoothed once by a moving average of three.

upwards of 23 inches long moved away from the tagging grounds during the summer of 1923 is not known for the reduced percentages of these larger fish taken as the summer and fall progressed may be due, at least in part, to the influx of smaller fish. Although the number of haddock taken on the four cruises made in 1923 is relatively small they were caught on many different days and are believed to be fairly representative of the small population that was present.

In 1924 evidence from length frequencies of the arrival of a new stock of haddock on Nantucket Shoals is not as striking as in 1923

for the dominant sizes showed relatively little change from July to October (Figure 7). However, the close resemblance of the July 1924 length distribution with that of October 1923, allowing for increased growth, gives some evidence that much of the population had remained localized during the interim.

In September 1924 it appears that some of the largest fish had moved away from the Orion Shoal-Rose and Crown section of Nan-

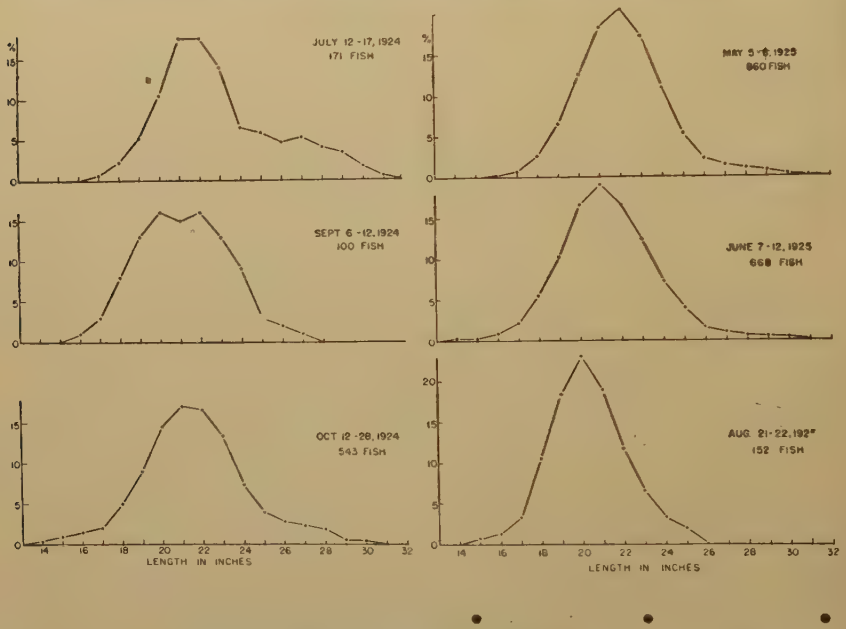


Figure 7. (Left.) Length frequency distributions by percentages of haddock caught by the *Halcyon* on Nantucket Shoals during 1924.

Figure 8. (Right.) Length frequency distributions by percentages of haddock caught by the *Halcyon* on Nantucket Shoals during 1925.

tucket Shoals and that possibly some smaller fish around 18–19 inches long had moved in. While in October the increased catch of haddock, per unit of effort, indicates that a fresh stock of fish, of about the same sizes as were present in September, had appeared on the tagging grounds.

In 1925 the frequency distributions again give evidence that many of the haddock present the previous fall (October 1924) remained over the winter, and that the large stock of fish then present had actually increased in numbers by May (Figure 8). In June haddock were

still plentiful but a shifting of the peak from 22 inches (in May) to 21 inches indicates that some of the larger haddock were leaving and this is still more apparent in August when the peak further retarded to 20 inches and haddock were found to be less abundant. There is little doubt but what most of the haddock present in the spring and summer had moved away by fall for in October 1925 they were found to be relatively scarce (Table II).

TABLE II—Catches of Haddock Made on Nantucket Shoals, in the Vicinity of Orion Shoal and Rose and Crown Buoys, by the *Halcyon* and the *Albatross II*, from 1923 to 1931

Year	Date	Hours of fishing	Number of haddock	Number per hour of fishing
1923	*April 19–May 4*	34	0	0.0
	May 23–28	48	48	1.0
	June 22–28	46	99	2.1
	August 16–23	42	36	0.8
	September 5–11	58	49	0.8
	October 3–8	41	129	3.0
	October 14–17	30	49	1.6
1924	*July 13–17*	47	177	3.8
	September 6–12	34	99	2.9
	October 16–22	29	174	6.0
	October 25–28	35	328	9.4
1925	*May 5–8*	35	859	24.5
	June 7–12	33	674	20.4
	August 20–25	31	236	7.6
	October 1–6	28	4	0.2
	October 24–30	10	48	4.8
1926	*September 6–11*	49	114	2.3
1927	*May 4–7*	17	17	1.0
	June 17–25	38	198	5.2
	September 1–3	29	194	6.7
	October 14–17	27	7	0.3
1928	*July 14–21*	35	186	5.3
	October 24–29	36	32	0.9
1929	*June 10–14*	33	119	3.6
1930	*August 18–20*	18	80	4.4
	October 18–22	27	5	0.2
1931	*August 10–11*	7	8	1.1

The catches of haddock per hour's fishing, as given in Table II, are based on factors which remained reasonably constant throughout. Thus the average number of hand lines which fished was about 6, the bait frozen herring, and the locality of fishing well marked by the proximity to Orion Shoal and Rose and Crown buoys. The relative abundance of haddock as found on these grounds (which are primarily cod-fishing areas) may not obtain for other parts of Nantucket Shoals, especially to the eastward where haddock are more plentiful. Nevertheless the comparisons given reveal periods of scarcity and abundance (as in May 1923 and May 1925) and show

that schools of haddock do move on to and away from this part of the Shoals.

CAPE COD TO BOOTHBAY HARBOR

Stellwagen Bank

On Stellwagen Bank, situated between Provincetown and Cape Ann, 192 haddock were tagged between the years 1924 to 1928. Four of these were reported recaptured, one on Platts Bank to the northeastward (506 days), one on Georges Bank to the southeastward (170 days), one off Chatham to the southward (101 days), and one which lacked sufficient data (Figure 3). These few records suggest that such Stellwagen haddock as migrate scatter in various directions.

Boone Island to Boothbay Harbor, Maine

Between Boone Island and Boothbay Harbor, a distance of about 60 miles, 1104 haddock were tagged between 1924 and 1930. Of these only 11 were reported recaptured, 2 of them locally (33–48 days) and 9 in distant localities (84–503 days[5]) (Figure 4). All but one of these latter were recaught to the southward of the tagging grounds to as far as Massachusetts Bay and Georges Bank. The remaining fish was caught to the eastward, off Prospect Harbor, 60 miles from Boothbay Harbor. Thus the indications are that most of the haddock found in the shore waters of this part of the coast of Maine are temporary residents and that, of those which move away, more of them go southward than go eastward. If an appreciable number of tagged fish had gone eastward, recaptures could have been expected off Mt. Desert and around the mouth of the Bay of Fundy where a rather intensive fishery is carried on.

MT. DESERT REGION

Off Mt. Desert 4174 haddock were tagged from 1924 to 1932, most of them within a few miles of Bakers Island and Little Duck Island and off Petit Manan, in a depth of 10 to 15 fathoms, and a few at Mt. Desert Rock in 15 to 25 fathoms. The recaptures reported totaled 144 fish, or 3.45 per cent, of which 114 fish, or 2.73 per cent were taken locally (average 244 days), 25 fish, or 0.60 per cent in distant localities (average 299 days) and 5 fish, or 0.12 per cent, lacked sufficient data. Compared with the 1.24 per cent returns from the Nantucket Shoals tagging, the larger yield from the Mt. Desert experi-

[5] One fish was reported recaptured nearly three years after tagging but there is some doubt as to the accuracy of this record.

ment was probably due to the intensive fishing in that vicinity from May to October.

The returns would have been much larger had a more efficient tag been used throughout the experiments. For example, in the experiments of 1924–1930 when metal tags were used, 4078 fish produced 131 recaptures, or 3.2 per cent, whereas from the 1930–1932 experiments when celluloid disk tags were attached to the tail and some fish carried celluloid belly tags, the 96 fish thus marked yielded a return of 12.5 per cent.

The relatively large number of recaptures resulting from the Mt. Desert tagging experiments shows rather definitely that a large proportion of the haddock present there at a given time remain localized for from six months to a year, or more. The elapsed days for the local recaptures are as follows: 46 were taken in 12–100 days; 7 in 101–200 days; 21 in 201–300 days; 19 in 301–400 days; 18 in 401–500 days; and 13 in 500 days or more. Of these the tagging vessels *Halcyon* and *Albatross II* recaught seven fish 49 to 62 days later, and one fish 309 days later. The records of some of the haddock at liberty for many days are not dependable as fishermen sometimes retained tags long after the capture, the date of which was forgotten. However, the haddock longest at liberty, 3 years and 253 days, carried a belly tag and its record is probably correct. The small number of recaptures made 101–200 days after tagging, as compared with the number reported 201–400 days later, is due to the fact that very little local fishing is done between December and March hence tagged haddock not caught locally by late fall would not be taken until at least the following spring. The experience of the fishery shows that in winter haddock move away from the grounds close to shore so it is probable that fish tagged one year and recaptured in the same vicinity the next made at least a short migration during the interim.

Localization of haddock off Mt. Desert is also indicated from the length frequency distributions of fish caught by the *Halcyon* and the *Albatross II*. Thus in the largest series of closely spaced experiments, from August 1924 to October 1925 the same body of haddock appears to have remained throughout (Figure 9). For example, in the September 29–October 5, 1924 cruise, the length distribution was essentially the same as in August, the slight change being what one might expect from two months growth, while in the summer and fall of 1925 the shifting of the dominant size group to 19–22 inches might also be attributed to the increased growth of the 17–19 inch August–October 1924 fish.

A slight retrogression of the dominant sizes occurred from May–

July to September, but they advanced again in October which was almost identical to July. This may have been caused by the arrival during the summer of a fresh stock of the smaller sizes or by the departure from the tagging grounds of more of the larger haddock than of the smaller. Furthermore, catches made by the local fishery probably accounted for more large fish than small for our experience has

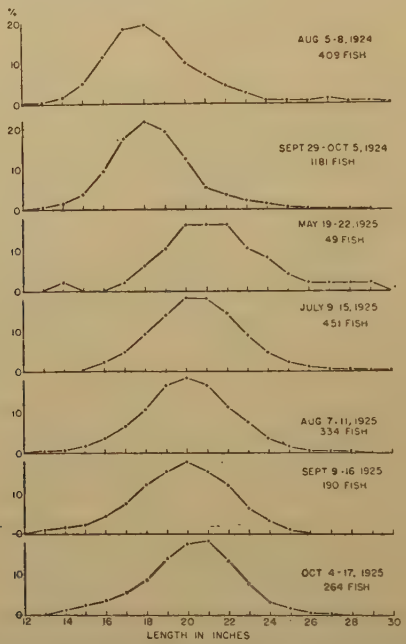


Figure 9. Length frequency distributions by percentages of haddock caught by the *Halcyon* off Mt. Desert, Maine, on cruises made from August 1924 to October 1925.

been that below a length of 17 inches hook and line fishing does not adequately sample the catch (Figure 1).

In localities distant from Mt. Desert 25 tagged haddock were reported recaptured (Figure 5). While this is a relatively small number compared with that taken locally (114) the chances of recapture after moving away must be much less than if the fish remained within the area of the intensively fished grounds close to shore. It appears quite certain that some of the emigrant tagged fish moved to localities where little or no fishing was done during the subsequent 6 or 12 months by which time many of them would have lost their tags. What proportion of the shore haddock emigrate within a year's time

cannot be determined from these experiments, but it is probably safe to say that at least one-half of them do so.

As 22 of the Mt. Desert migrant tagged haddock were recaptured between Penobscot Bay and the South Channel-Georges Bank region, compared with only 3 reported from the eastward (Grand Manan) the migratory tendency of haddock living along the coast of eastern Maine appears to be decidedly to the southward. Certainly if many of the fish went eastward more recaptures would have resulted from the intensive fishing around the mouth of the Bay of Fundy and southern Nova Scotia. The recapture localities and elapsed days at liberty for these migrants are as follows:

Grand Manan, N. B.	288	Monhegan	988
Dipper Harbor, N. B.	304	Jeffreys Ledge	260, 442
Bay of Fundy	265	Platts Bank	93, 119, 379
Penobscot Bay	280, 296	Cape Porpoise	259
Damarisicove	318	Manchester	311
Portland	126, 434, 715	South Channel	130, 134, 224, 520
		Georges Bank	154, 226, 308, 326.

OFFSHORE GROUNDS

Negligible results were obtained from the haddock tagged on the offshore grounds. As these experiments were made primarily for cod, haddock were caught only incidentally and the total number tagged was relatively small. Furthermore, it is probable that many of the fish taken in deep water (20 to 50 fathoms), all caught with hand lines, failed to survive after liberation. And the offshore grounds, especially Georges Bank and Browns Bank, are relatively so large as compared with the shore grounds, that individual fish would seem to stand a better chance to elude capture there. The method of catching and handling fish on Georges Bank was no doubt partly responsible for the small number of tagged haddock reported from there (a total of 8 fish,—2 from Nantucket Shoals, 1 from Stellwagen Bank, 1 from off Boothbay Harbor, and 4 from Mt. Desert). Most of the commercial catch was taken with otter trawls so that a marked fish would stand very little chance of being detected until it reached the cleaning board, by which time its tag would likely be dislodged and, later, probably washed down the scuppers unnoticed.

Platts Bank, a small isolated plateau located about 30 miles southeast of Portland lightship might have provided some interesting results had the tagging experiments there yielded sufficient returns. However, from 1924 to 1930, out of 237 marked haddock only one was reported recaptured, a fish liberated in August 1928 and recaught in June 1929 off Rockland about 60 miles inshore to the northeast.

On Browns Bank 747 haddock were tagged from 1927 to 1930, and only one was reported recaptured. This fish was tagged in August 1930 and recaptured on Browns Bank in February 1931.

No recaptures were reported from other haddock tagged offshore from 1926 to 1930, as follows: Georges Bank, 246; Cashes Ledge, 174; and Jeffreys Ledge, 53. Also, in 1928, 25 haddock were tagged off Cape Sable with no recaptures reported.

AGE AND RATE OF GROWTH

During the course of these tagging experiments relatively little effort and study was given to the problem of the haddock's age and rate of growth. However, numbers of very small haddock were measured from time to time and enough records were obtained to

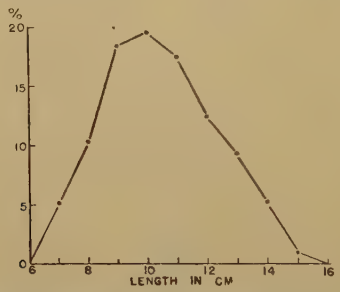


Figure 10. Length frequency distribution by percentages of 98 juvenile haddock taken on Georges Bank in miscellaneous catches from August 13 to September 25, 1925 to 1929.

give some indication of the size attained during the first year or two after spawning. Some of these young fish were taken on bottom in otter trawls, off bottom in plankton nets, and from cod and pollock stomachs, by the *Halcyon* and the *Albatross II*. Others were caught by otter trawlers operating on Georges Bank.

The smallest fish were caught from August 13 to September 15, during the years 1925 to 1929, when miscellaneous catches of 98 juvenile haddock were made on Georges Bank. These fish ranged in length from 6 to 15 cm. (Figure 10) most of them (76 fish) between 8 and 12 cm. (3⅛–4¾ inches).

From February 9 to March 1, 1930 a large sample of young haddock (500 fish) was taken on Georges Bank by the trawler *Widgeon* and measured by Mr. Robert Smith, the wireless operator (Figure 11). They fall into two distinct age groups. The smaller fish (12–21 cm.) could easily have passed through the meshes of the net, but so many became trapped among the great mass of large fish landed on deck

that the samples can be considered fairly representative. These haddock very likely were spawned the previous spring and they had attained an average size of around 15–16 cm., or 6–6¼ inches, at about 10 months of age. Forty-eight fish taken on Georges Bank the following March 11–April 11 were of the same size range, with the dominant group also at 15–16 cm. This average length appears to obtain for fish about one year of age.

The 25 to 35 cm. haddock (165 fish) represent the smallest scrod that were taken (February 9 to March 1) and, while they are there-

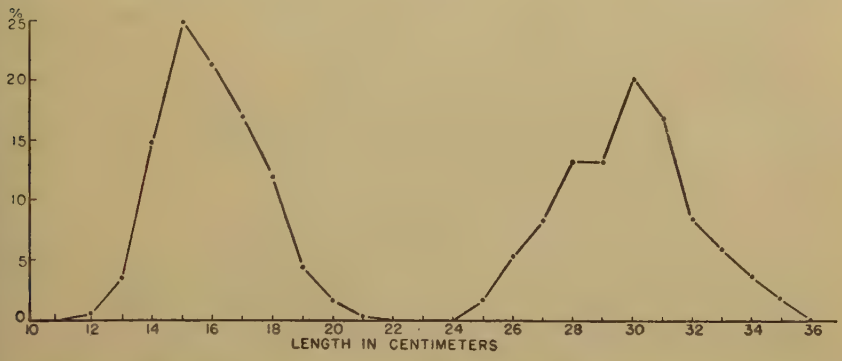


Figure 11. Length frequency distribution by percentages of young haddock caught on Georges Bank from February 9 to March 1, 1930 by the trawler *Widgeon*. Each curve represents 100 per cent.

fore not an unselected sample, nevertheless the dominant group at 30 cm. (12 inches) suggests that Georges Bank haddock are approximately this size when about two years old. A subsequent sample of 70 fish, taken March 11, also averaged about 30 cm. in length.

No attempt was made to study the haddock's rate of growth beyond the second year although scale samples were obtained from virtually all the fish that were tagged after 1923. And very few reliable records were received from recaptured fish, for in nearly every instance no remeasurement was made by the fishermen or, if the length was mentioned, it appeared to be incorrect. Following are the records of the ten fish recaptured by the *Halcyon* and the *Albatross II* arranged in the order of days at liberty:

Tag No.	Date Tagged	Length	Date Recaptured	Length	Days at Liberty
33018	June 26, 1925	22¼	August 13, 1925	22½	48
30843	May 21, 1925	21	July 9, 1925	21½	49
20474	August 8, 1924	18¼	October 3, 1924	18½	56
20237	August 8, 1924	17¼	October 5, 1924	17¾	58
19853	August 7, 1924	16¾	October 4, 1924	17	58
19797	August 5, 1924	18½	October 5, 1924	18¾	61
34139	July 14, 1925	26½	September 14, 1925	26½	62
24899	October 3, 1924	16¾	July 9, 1925	18¾	279
53459	September 1, 1927	24¾	July 19, 1928	25¼	321
18854	July 16, 1924	23¼	June 11, 1925	24	330

Thus six fish 16¾–22¼ inch long had gained 0.25 to 0.50 inch in about two months, one fish 26½ inches long showed no increase in that period, one 16¾ inch fish gained 2 inches in nine months and two fish 23¼–24¾ inches long increased 0.50 to 0.75 inch in eleven months. Admittedly these fragmentary data throw but little light on the growth of adult haddock.

SUMMARY

In the Nantucket Shoals region a substantial part of the adult haddock population remains more or less localized for periods up to from 6 to 12 months, or more. Those haddock which migrate or wander away from the Nantucket region appear to scatter to the northward of Cape Cod, alongshore as far as the Bay of Fundy, and also eastward to Georges Bank.

Haddock present along the southwestern coast of Maine, from Boone Island to Boothbay Harbor, appear to migrate to the southward to Massachusetts Bay and to as far as Georges Bank.

Along the northeastern coast of Maine, in the vicinity of Mt. Desert, many of the haddock remain localized for periods up to from 6 to 12 months, or more. Those fish which migrate go chiefly to the southward to as far as the South Channel region and Georges Bank. A small part of the population moves eastward to the Bay of Fundy.

Thus within the Gulf of Maine southern haddock tend to move northward and northern haddock tend to move southward.

The number of tagged fish recaught was too small to reveal any seasonal migrations.

No information was obtained on the migratory habits of haddock living on Browns Bank and Georges Bank and on certain smaller grounds offshore as only two recaptures resulted from these experiments.

According to the length frequency distributions of young fish taken on Georges Bank, haddock in that region are about 6 inches long at one year of age and about 12 inches long at two years.

LIST OF RECAPTURES OF TAGGED HADDOCK

Explanation: Each entry gives the data for one fish, the first item of the entry giving the date of release, the second item the place of recapture, the third item the date of recapture, and the fourth item (in italics) the number of days at liberty. VI.24.1923, South Channel VI.23.1924, *365;* should thus read: released June 24, 1923, recaptured in South Channel June 23, 1924 after 365 days at liberty.

Fish released on Nantucket Shoals

VI.24.1923, South Channel VI.23.1924, *365;* VI.24.1923, off Chatham IV.23.1925, *666;* VI.25.1923, Unknown II.?.1924; IX.10.1923, Unknown IV.?.1924; X.4.1923, Nantucket Shoals VII.2.1924, *271;* X.16.1923, Nantucket Shoals X.28.1924, *377;* VII.16.1924, Nantucket Shoals VI.11.1925, *330;* IX.12.1924, Stellwagen Bank XI.20.1924, *69;* X.18.1924, S. E. Georges Bank II.9.1925, *114;* X.18.1924, Stellwagen Bank VII.30.1925, *286;* X.27.1924, Unknown III.?.1925; X.27.1924, South Channel VIII.17.1925, *294;* X.27.1924, South Channel II.16.1925, *112;* X.27.1924, South Channel VIII.6.1925, *283;* X.28.1924, Nantucket Shoals VII.31.1925, *276;* X.28.1924, South Channel VIII.18.1925, *294;* V.6.1925, South Channel VIII.27.1925, *113;* V.6.1925, South Channel VIII.2.1925, *88;* V.6.1925, Nantucket Shoals VIII.27.1925, *113;* V.6.1925, South Channel, VI.30.1926, *420;* V.6.1925, Stellwagen Bank VIII.8.1925, *94;* V.6.1925, Off Chatham VIII.9.1925, *95;* V.7.1925, South Channel X.26.1925, *172;* V.7.1925, Off Chatham VIII.25.1925, *112;* V.7.1925, South Channel VII.25.1925, *79;* V.8.1925, Off Chatham IX.12.1925, *127;* V.8.1925, Nantucket Shoals VIII.16.1926, *465;* VI.7.1925, Salisbury Beach, Mass. VIII.5.1926, *424;* VI.7.1925, South Channel V.13.1926, *340;* V.8.1925, South Channel VIII.27.1925, *81;* VI.10.1925, Platts Bank X.19.1925, *131;* VI.11.1925, Off Chatham VII.1.1925, *20;* VIII.20.1925, Head Harbor, N. B. VI.8.1926, *292;* X.6.1925, South Channel XI.12.1925, *37;* V.3.1927, South Channel X.11.1927, *161;* IX.1.1927, Nantucket Shoals VII.19.1928, *321**; VII.13.1928, Newburyport, Mass. XII.18.1928, *158;* VII.15.1928, South Channel X.6.1928, *83;* VII.15.1928, South Channel X.15.1929, *457;* VII.15.1928, Unknown 1933; VII.19.1928, Unknown 1930; VII.20.1928, Off Chatham VII.16.1929, *361;* VII.20.1928, Off Chatham X.21.1928, *93;* VII.20.1928, Nantucket Shoals X.14.1928, *86;* VI.11.1929, South Channel X.15.1929, *127;* VI.12.1929, Grand Manan, N. B. IX.25.1929, *105;* VI.12.1929, South Channel II.15.1930, *246;* X.22.1930, Georges Bank III.6.1931, *156.*

Fish released in South Channel

VI.13.1929, South Channel VIII.22.1929, *70.*

Fish released on Stellwagen Bank

X.16.1925, Unknown VII.?.1926; X.16.1925, Off Chatham I.25.1926, *101;* X.16.1925, Platts Bank III.6.1927, *506;* X.16.1928, E. Georges Bank IV.7.1929, *170.*

Fish released at Boone Island

X.12.1924, Boston Lightship IV.22.1925, *192.*

Fish released off Portland

VI.26.1925, Off Portland VIII.13.1925, *48**; VII.22.1925, Off Portland VIII.24.1925, *33.*

Fish released at Seguin

VII.31.1924, Off Portland VI.8.1925, *312.*

Fish released off Boothbay Harbor

IX.22.1924, Portland Market IV.15.1927, *935;* IX.22.1924, Platts Bank XII.15.1924, *84;* IX.24.1924, Prospect Harbor, Me. IX.17.1925, *358;* IX.24.1924, Platts Bank II.9.1926, *503;* IX.25.1924, Portland I.12.1925, *109;* VII.18.1925, Georges Bank X.10.1926, *449;* IX.4.1925, Stellwagen Bank III.15.1926, *192.*

Fish released at Mt. Desert

VIII.5.1924, Mt. Desert X.5.1924, *60**; VIII.7.1924, Mt. Desert X.4.1924, *57**; VIII.7.1924, Mt. Desert X.?.1924, *60;* VIII.8.1924, Mt. Desert IX.27.1924, *50;* VIII.8.1924, South Channel III.20.1925, *224;* VIII.8.1924, Mt. Desert "Fall".1924, *60;* VIII.8.1924, Mt. Desert X.5.1924, *58**; VIII.8.1924, Mt. Desert "X-XI".1925, *400;* VIII.8.1924, Mt. Desert "Fall".1924, *60;* VIII.8.1924, Mt. Desert "V-VI".1925, *300;* VIII.8.1924, Mt. Desert X.3.1924, *56**; VIII.8.1924, Mt. Desert "Fall".1924, *60;*

VIII.8.1924, Mt. Desert VII.?.1925, *330;* VIII.10.1924, Mt. Desert "VI-VII".1925, *330;* VIII.10.1924, Mt. Desert "V-VI".1925, *500;* IX.30.1924, Mt. Desert "VI-VII".1925, *300;* IX.30.1924, Mt. Desert "V-VI".1925, *270;* IX.30.1924, Dipper Harbor, N. B. VIII.1.1925, *304;* IX.30.1924, Mt. Desert Rock IV.14.1925, *196;* IX.30.1924, Mt. Desert "Fall".1925, *65;* IX.30.1924, Mt. Desert "Fall".1925, *65;* IX.30.1924, Mt. Desert "V-VI".1925, *240;* X.2.1924, Mt. Desert VIII.10.1925, *312;* X.2.1924, Platts Bank I.29.1925, *119;* X.2.1924, Mt. Desert "X-XI".1925, *390;* X.2.1924, Mt. Desert V.19.1927, *949;* X.2.1924, Mt. Desert X.?.1924, *30;* X.3.1924, Mt. Desert "Fall".1925, *360;* X.3.1924, Mt. Desert "Fall".1924, *30;* X.3.1924, Jeffreys Ledge XII.22.1925, *442;* X.3.1924, Mt. Desert VII.9.1925, *309**; X.3.1924, Mt. Desert IX.?.1925, *350;* X.3.1924, Mt. Desert VIII.?.1926, *660;* X.3.1924, Mt. Desert VIII.?.1925, *300;* X.4.1924, Mt. Desert "VI-VII".1925, *270;* X.4.1924, Mt. Desert "Spring".1927, *940*?; X.4.1924, Mt. Desert XII.8.1924, *65;* X.5.1924, Platts Bank X.19.1925, *379;* X.5.1924, Grand Manan, N. B. VII.21.1925, *288;* V.19.1925, Mt. Desert VII.?.1925, *57;* V.19.1925, Mt. Desert "VII-VIII".1925, *70;* V.19.1925, Mt. Desert "VI-VII".1925, *60;* V.20.1925, Mt. Desert VII.?.1925, *60;* V.21.1925, Mt. Desert VII.9.1925, *49**; V.22.1925, Mt. Desert VIII.?.1925, *80;* VII.9.1925, Mt. Desert VIII.31.1925, *53;* VII.9.1925, Mt. Desert VII.?.1925, *20;* VII.9.1925, Portland? I.8.1927, *715;* VII.9.1925, Mt. Desert VII.21.1925, *12;* VII.10.1925, Mt. Desert "X-XI".1926, *465;* VII.10.1925, Mt. Desert "X-XI".1926, *465;* VII.10.1925, Mt. Desert "Summer" 1926, *365;* VII.10.1925, Mt. Desert XII.?.1925, *170;* VII.10.1925, Mt. Desert XI.30.1925, *143;* VII.10.1925, Mt. Desert VIII.8.1925, *29;* VII.11.1925, Mt. Desert, VII.25.1925, *14;* VII.11.1925, Mt. Desert "X-XI".1925, *110;* VII.11.1925, Mt. Desert VIII.20.1925, *40;* VII.13.1925, Unknown II.?.1926; VII.14.1925, Mt. Desert IX.14.1925, *62**; VII.14.1925, Mt. Desert "Spring" 1926, *270;* VII.14.1925, Mt. Desert VIII.20.1925, *37;* VII.15.1925, Mt. Desert "X-XI".1925, *110;* VII.15.1925, Mt. Desert X.?.1926, *470;* VII.15.1925, Mt. Desert "VII-VIII".1925, *30;* VII.15.1925, Mt. Desert X.?.1926, *470;* VII.15.1925, Mt. Desert "Fall" 1927, *790;* VII.15.1925, Mt. Desert "VII-VIII" 1925, *30;* VII.15.1925, Mt. Desert IX.?.1926, *400;* VII.15.1926, Mt. Desert 1928?; VIII.7.1925, Mt. Desert "Spring-Summer" 1927, *700;* VIII.8.1925, Platts Bank XI.9.1925, *93;* VIII.8.1925, Mt. Desert, *360;* VIII.8.1925, Mt. Desert "IV-V". 1926, *270;* VIII.8.1925, Mt. Desert "Spring-Summer" 1927, *720;* VIII.8.1925, Mt. Desert "Spring-Summer" 1927, *720;* VIII.8.1925, Mt. Desert "Spring-Summer" 1927, *720;* VIII.8.1925, Mt. Desert "Spring-Summer" 1927, *720;* VIII.8.1925, Mt. Desert "X-XI".1925, *70;* VIII.11.1925, Mt. Desert "Summer" 1926, *360;* X.14.1925, Georges Bank III.17.1926, *154;* X.15.1925, Mt. Desert XII.23.1926, *434;* X.15.1925, Mt. Desert "IV-V".1926, *180;* X.16.1925, Rockland Bay VIII.8.1926, *296;* X.16.1925, Mt. Desert "X-XI". 1926, *365;* X.16.1925, Mt. Desert "X-XI".1926, *365;* X.16.1925, Mt. Desert "Summer" 1926, *270;* X.16.1925, Mt. Desert "Summer" 1926, *270;* VIII.21.1926, Mt. Desert IX.?.1926, *30;* VIII.21.1926, Mt. Desert "Spring" 1927, *210;* VIII.23.1926, Mt. Desert VII.20.1927, *332;* VIII.23.1926, Bay of Fundy V.15.1928, *265;* IX.26.1926, Mt. Desert "Fall" 1927, *365;* IX.27.1926, Mt. Desert X.15.1926, *18;* IX.27.1926, South Channel II.4.1927, *150;* IX.27.1926, Mt. Desert "Spring-Summer" 1927, *270;* IX.27.1926, Mt. Desert "Spring-Summer" 1927, *270;* IX.27.1926, Mt. Desert "X-XI".1926, *40;* IX.27.1926, Mt. Desert "X-XI".1926, *40;* IX.27.1926, Mt. Desert "XII-I".1927, *100;* IX.28.1926, Mt. Desert "Spring-Summer" 1927, *210;* IX.28.1926, Mt. Desert "Spring-Summer" 1928, *600;* IX.29.1926, Mt. Desert "X-XI".1926, *40;* IX.29.1926, Mt. Desert "Spring-Summer" 1927, *210;* VII.15.1927, Mt. Desert X.1.1928, *442;* VII.15.1927, Mt. Desert VIII.15.1928, *395;* VII.15.1927, Unknown 1928; VII.18.1927, Mt. Desert "Spring-Summer" 1928, *300;* VII.18.1927, Unknown III.?.1928; VII.18.1927, Mt. Desert 1928, *250;* VII.18.1927, Manchester, Mass. V.25.1928, *311;* VII.18.1927, Mt. Desert VIII.?.1927, *30;* VII.19.1927, Mt. Desert "Spring-Summer" 1928, *270;* VII.19.1927, Cape Porpoise, Me. IV.4.1928, *259;* VII.19.1927, Mt. Desert X.28.1927, *101;* VII.20.1927, Mt. Desert VI.8.1928, *323;* X.31.1927, Mt. Desert "X-XI".1929, *730;* VIII.19.1928, Mt. Desert "Fall" 1928, *60;* VIII.20.1928, Mt. Desert "X-XI".1929, *440;* X.20.1928, Penobscot Bay V.27.1929, *280;* VIII.20.1928, Mt. Desert IX.25.1929, *401;* VIII.20.1928, Jeffreys Ledge V.7.1929, *260;* VIII.20.1928, Mt. Desert "Spring" 1930, *620;* VIII.21.1928, S. W. Georges Bank VI.24.1929, *308;* VIII.22.1928, Mt. Desert "Fall" 1928, *60;* VIII.22.1928, Mt. Desert VIII.19.1929, *362;* VIII.22.1928, Mt. Desert X.1.1928, *40;* VII.19.1930, Mt. Desert VIII.?.1930, *30;* VII.19.1930, Mt. Desert IX.?.1930, *60;* X.4.1930, Mt. Desert VI.14.1934, *1348;* X.9.1930, South Channel III.?.1932, *520;* VI.23.1930, Lat. 41° 20′ N.; Long. 67° 10′ W., *326;* VI.23.1930, Damariscove, Me. V.7.1932, *318;* VI.23.1931, Mt. Desert IV.?.1932, *300;* VI.24.1931, Mt. Desert VIII.25.1931, *62;* VI.24.1931, Georges Bank III.9.1932, *226;* VI.24.1931, Mt. Desert VII.16.1931, *22;* VI.27.1931, Mt. Desert VII.23.1931, *26;* VI.27.1931, Boston Fish Pier V.?.1932; VI.27.1931, Mt. Desert I.18.1932, *203;* VII.16.1932, 20 mi. S. E. Monhegan III.31.1925, *988.*

Fish released at Mt. Desert Rock

IX.15.1925, Off Portland XI.24.1926, *434;* IX.15.1925, Unknown VI.?.1929; X.27.1927, South Channel III.10.1928, *134;* X.27.1927, Off Portland III.2.1928, *126.*

Fish released on Platts Bank

VIII.18.1928, Rockland Bay, Me. VI.10.1929, *296.*

Fish released on Browns Bank

VIII.2.1930, Browns Bank II.19.1931, *201.*

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*Recaptured by the *Halcyon* or the *Albatross II.*

SIGMA-T SURFACES IN THE ATLANTIC OCEAN

By

R. B. MONTGOMERY AND M. J. POLLAK

New York University and *Woods Hole Oceanographic Institution*[1]

Wüst (1936) shows the distribution of σ_t on a number of surfaces of constant depth in the Atlantic Ocean, based on about seventy expeditions over a period of sixty years. For most purposes there is more significance in the configuration of σ_t-surfaces, since these are approximately the surfaces of isentropic flow in the absence of vertical mixing. Such preliminary charts have been obtained directly from those of Wüst without recourse to the original data. Wüst's charts give the lines of intersection of σ_t-surfaces with surfaces of constant depth, arranged according to depth; the charts presented in Figures 12–17 are tracings of the same lines, but arranged according to σ_t.

In the construction of surfaces $\sigma_t = 26.5$ g l^{-1} and $\sigma_t = 27.0$ g l^{-1} Montgomery's (1938) σ_t-charts of the southern North Atlantic also were utilized. The contours in all the charts were simplified to some extent. A number of inconsistencies in the original lines giving instability have been eliminated. Due to the greater spacing of the deeper surfaces of constant depth a contour interval of 500 meters was used on the surfaces $\sigma_t = 27.6$ g l^{-1} and $\sigma_t = 27.8$ g l^{-1} instead of the 200-meter interval used on the surfaces of lower σ_t-values.

The boundaries of σ_t-surfaces are their intersections with the continental slopes, indicated by broken lines, and with the sea surface. The mean positions of the 0-contour were taken from Böhnecke's (1936, Beilage L) chart of the mean density of the sea surface. The polar and equatorial extremes of the intersections with the sea surface, indicated by dotted lines, have been estimated from Böhnecke's monthly and quarterly charts.

[1] Contribution No. 305 from Woods Hole Oceanographic Institution.

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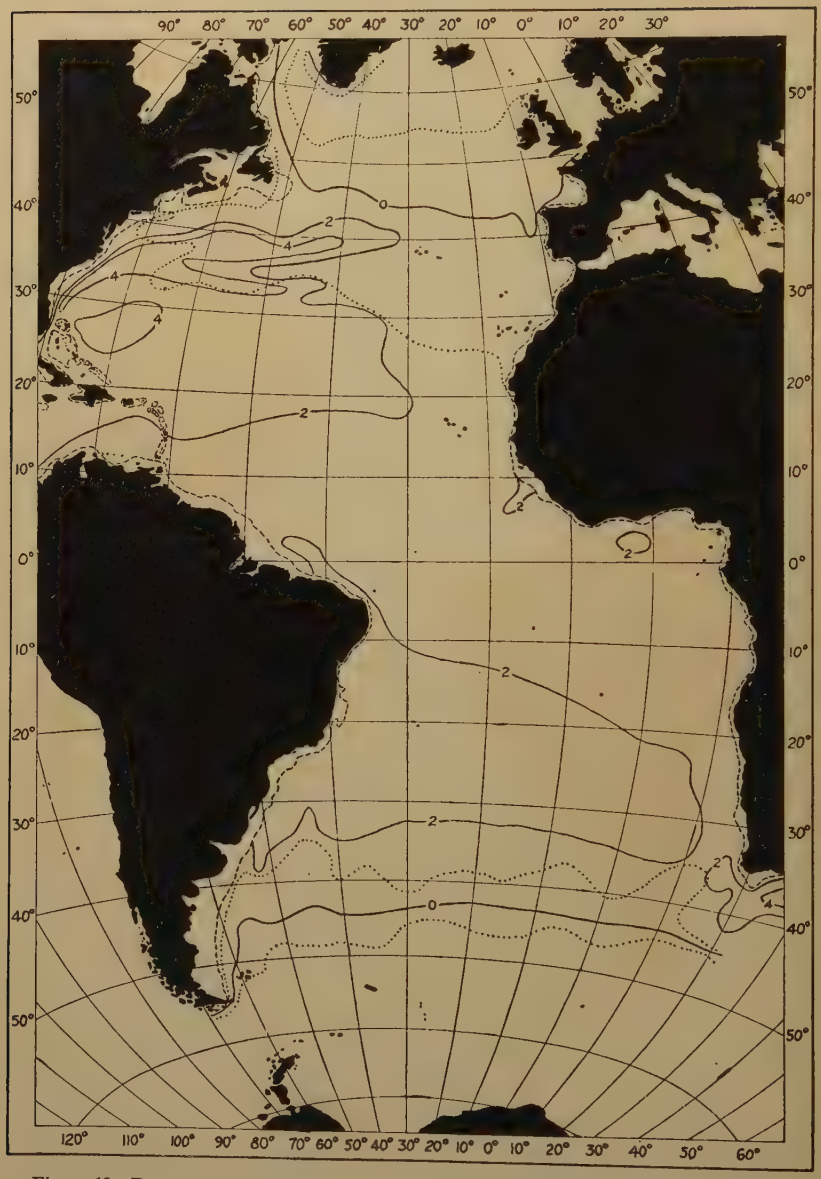

Figure 12. Depth in hektometers of $\sigma_t = 26.5$ g l^{-1}. Azimuthal equal-area projection.

Figure 13. Depth in hektometers of $\sigma_t = 27.0$ g l^{-1}. Azimuthal equal-area projection.

Figure 14. Depth in hektometers of $\sigma_t = 27.2$ g l^{-1}. Azimuthal equal-area projection.

Figure 15. Depth in hektometers of $\sigma_t = 27.4$ g l^{-1}. Azimuthal equal-area projection.

Figure 16. Depth in hektometers of $\sigma_t = 27.6$ g l^{-1}. Azimuthal equal-area projection.

Figure 17. Depth in hektometers of $\sigma_t = 27.8$ g l^{-1}. Azimuthal equal-area projection.

THE BROMINE-CHLORINITY RATIO OF SEA WATER

By

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Dittmar (3) first showed that a constant ratio apparently exists between the bromides and the total halides in the waters of the several oceans. Some years later Winkler (13) determined the bromine in samples of sea water, and the bromine-chlorinity ratio calculated from his results showed remarkable agreement but were slightly higher than those of Dittmar. The ratio was obtained from the expression

$$\frac{\text{Grams Br per kilo}}{\text{‰ Cl}} = K$$

where ‰ Cl is the chlorinity of the sea water, defined until recently as the total weight in grams of halides calculated as chlorides in a kilogram of sea water.

The bromine content of sea water was determined by Dittmar using a method of fractional precipitation while Winkler oxidized the bromides and aspirated the liberated bromine. The former's value for K was 0.0034_1 and the latter's, 0.0034_7. Thus the establishment of the constancy of this ratio permits the estimation of bromine in sea water from the chlorinity by a simple calculation for practical purposes.

In recent investigations in the Japan Sea, Vasil'ev (12) reported that the bromine content of these waters was about 2 mg. less per kilogram of sea water than expected from the ratio of Winkler and Dittmar. On the other hand, Ratmanoff (9) showed that the waters of Bering Strait and Bering Sea were considerably higher in concentrations of bromine than those of the Pacific. Using his own data and those of Vasil'ev, he demonstrated that the bromine-chlorinity ratio decreases with a decrease in latitude. From studies made by one of us of the circulation of waters of the Northeast Pacific and Bering Sea, the conclusions of Vasil'ev and Ratmanoff were doubted. Recently Miyake (8) has investigated the bromine content of waters of the western Pacific and obtained a ratio approximating that of Dittmar and Winkler. However he used the colorimetric method of Dibdin and Cooper (2) which he claimed may not be as accurate as other methods.

In view of the above it was decided to make a study of the bromine content of waters of the Northeast Pacific, Bering Sea, and Bering Strait.

SOURCES OF SAMPLES STUDIED

The authors collected a number of samples at different depths and localities in Bering Strait and Bering Sea from the U. S. Coast Guard Cutter *Northland* and in the Northeast Pacific and Puget Sound regions from the University of Washington research ship *Catalyst* (11). All samples were secured with the Nansen-Knudsen bottles equipped with standard protected reversing thermometers. Samples of water from the Antarctic Ocean were secured by the British-Australian-New Zealand Antarctic Research Expedition, the chlorinities being determined by A. Howard, chemist of the expedition, and checked later by the authors.

METHODS OF ANALYSIS

Chlorinity. The chlorinity was determined by titration with silver nitrate solution using potassium chromate as an indicator (10). The silver nitrate solution was standardized against the standard sea water supplied by the Hydrographical Laboratories of Copenhagen.

Chlorinity of the standard water of Copenhagen is based upon the atomic weights of 1901 and by international agreement is used for all oceanographic investigations as a basis of chlorinity and salinity determinations. (It should not be used as a standard for other constituents of sea water.) Jacobsen and Knudsen (4) have shown that 58.99428 grams of atomic weight silver prepared especially by Otto Hönigschmid is equivalent to 19.3809 grams of chloride of the standard water. Chlorinity is now defined in respect to a unit weight of silver thus insuring a constancy of comparison of sea water throughout the world and independent of changes in atomic weight. A solution of sodium chloride containing 35.457 grams of chloride per kilogram of solution has a chlorinity of 35.441.

In the determination of the chlorinities reported here the maximum variation between extremes was 0.012‰ Cl with an average variation of ±0.004‰ Cl.

Bromine. The van der Meulen method described by Kolthoff and Yutzy (6) gave excellent results using sodium formate solution to reduce the excess of hypochlorite instead of hydrogen peroxide. The standard solutions required for the determination were prepared as follows:

Standard Potassium Bromide: 11.902 grams of pure potassium bromide free of chloride and prepared by thermal decomposition of pure potassium bromate were dissolved and diluted in a liter volumetric

flask. 100 ml. of this solution were pipetted and diluted to exactly one liter, and aliquots of this were used for standardizing the thiosulfate solution as described below. The actual molarity of the bromide solution was ascertained by taking aliquots, precipitating AgBr with a slight excess of silver nitrate, and weighing the precipitate.

Standard Sodium Thiosulfate Solution: 5 grams $Na_2S_2O_3 \cdot 5H_2O$ were dissolved in freshly boiled distilled water to give a volume of 2 liters. A few drops of CS_2 were placed in the solution to stabilize it. The solution was standardized as follows: 10 ml. of standard 0.0100 M potassium bromide solution were pipetted into a 600 ml. Erlenmeyer flask and diluted to 100 ml. 10 ml. NaH_2PO_4 solution were introduced and then 5 ml. NaOCl solution. The solution was then heated and boiled. When partially cooled, 5 ml. $H \cdot COONa$ solution were added. If the latter is introduced while the solution containing the excess hypochlorite is near the boiling point, the sudden evolution of CO_2 may give low results. After decomposition of the excess of hypochlorite, the solution was cooled and 10 ml. 6N H_2SO_4 were poured into the flask. 10 ml. of KI solution were then added and a drop of ammonium molybdate solution to catalyze the reaction between the bromate and the iodide ions. The liberated iodine was titrated with $Na_2S_2O_3$ using starch as an indicator and finally the normality of the solution calculated. Blank determinations were essential.

Determination of bromine in sea water was made by taking 100 ml. sample at 20° C. and treating it with the various reagents as described above. The excess hydroxide in the hypochlorite reagent caused precipitation of magnesium hydroxide and thus care was necessary to prevent bumping on boiling. The liberated iodine was titrated and the bromine content per liter of sea water calculated. From the chlorinity of sea water the density of each sample at 20° C. was ascertained from Knudsen's Hydrographical Tables (5), and the bromine content per kilogram of sea water was determined.

The method was found to be very reliable; the average variation between duplicates on 151 samples was about 0.25%.

DISCUSSION OF DATA

In Table I are given the temperatures *in situ,* chlorinities, grams of bromine per kilogram of sample, and the bromine-chlorinity ratios for waters secured from Bering Strait and northern Bering Sea. Much of this area is very shallow, and seldom do depths as great as 50 meters occur. A current flows continually in a general northerly direction from Bering Sea through Bering Strait into Chukchi Sea. The velocity of this current may be decreased or increased by wind

and during the winter influenced by ice. Table II contains similar data for the southern portion of Bering Sea, the samples having been taken in the vicinity of the Pribilof Islands and about 500 miles south of Bering Strait.

In cruising from Dutch Harbor to the Strait of Juan de Fuca surface samples were obtained hourly. Many of the samples taken in the Pacific in the area about 1000 miles east of Dutch Harbor and 250 miles west of the Strait of Juan de Fuca were analyzed for bromine. These data are given in Table III.

Two stations for water sampling were occupied beyond the continental shelf off the coast of the State of Washington, and samples were secured at various depths. Data from these two stations are given in Tables IV and V. Samples were likewise taken from various depths in the Strait of Juan de Fuca and at the entrance to Admiralty Inlet, Puget Sound, and the data are given in Tables VI and VII.

In Table VIII are given the results of determinations made on waters from the Antarctic Ocean. Because of the limited volume of samples available, 50 ml. of sea water were used instead of 100 ml.

For all the samples reported above, an average ratio of 0.0034_7 was obtained. Extremes of experimental errors could cause the ratio to vary from 0.0034_4 to 0.0035_1. However for the particular methods used, the average deviation from the mean should be less than ± 0.00002. The data show remarkable constancy for the bromine-chlorinity ratio but are not in agreement with observations and conclusions of Vasil'ev (12) and Ratmanoff (9). It is believed that this discrepancy is due to experimental errors of their method for determination of bromine. (The method of Deniges and Chelle (1) was used by them after a study of other procedures.)

Samples from the Antarctic gave an average bromine-chlorinity ratio that was slightly lower than that for northern waters. These samples were collected from eight to ten years prior to the others investigated, but they had been carefully stored in the original containers, doubly sealed, until analysis in 1940.

Lyman and Fleming (7) give Dittmar's ratio, corrected for atomic weights, as 0.00340. This value has been generally accepted. Dittmar's ratio published in 1884 for the oceans of the world is slightly less than that reported above while that of Winkler's for a limited number of samples is in much better agreement. Ratios calculated from data of other investigators vary from 0.0029 to 0.0044 but much of these data resulted from using methods of questionable or insufficient accuracy. Very decided variations are reported by individual investigators. Thus the 55 samples analyzed by Vasil'ev and Ratmanoff

show bromine-chlorinity ratios from 0.0032 to 0.0044 but the average of all of them is 0.0034_5.

Different methods of analysis for bromine will tend to give different results, the variations depending upon the accuracies and precisions of the method and the presence of interfering substances in the samples of sea water. Thus the presence of iodides and iodates would tend to give high results for bromine. These ions occur in sea water (7, 11) but in very small quantities. The iodides if present would be oxidized to iodates which would later react with the potassium iodide introduced during the determination. However the error would be about $+0.00001$ in the bromine-chlorinity ratio. In like manner any other substances present that would be oxidized by the hypochlorite giving compounds that would eventually liberate iodine from the potassium iodide, would increase the ratio. Bromates do not occur in sea water.

SUMMARY

1. The bromine-chlorinity ratios for waters of Bering Sea and the Northeast Pacific have been determined and found to be remarkably constant.

2. The bromine content of sea water when using the method described may be ascertained readily from the chlorinity of a sea water by multiplying by the factors 0.0034_7 to give the concentration as grams or 0.043_4 as milligram atoms.

TABLE I—Grams Bromine Per Kilo of Sea Water and the Bromine-Chlorinity Ratio Bering Strait and Northern Bering Sea. July 31 and August 1, 1937

Depth *Meters*	*Temp.* °C.	*Cl* °/oo	*Br* °/oo	*Br/Cl*	*Location*
10	3.91	17.67	0.0613	0.0034_7	Lat. 66° 04′ N., Long. 168° 14′ W.
50	3.04	17.75	.0616	34_7	
0	9.72	17.26	.0606	35_1	Lat. 65° 46′ N., Long. 168° 38′ W.
10	5.98	17.86	.0618	34_8	
0	12.79	15.82	.0551	34_8	Lat. 65° 22′ N., Long. 168° 15′ W.
0	13.58	14.56	.0507	34_8	Lat. 64° 40′ N., Long. 167° 07′ W.
20	2.30	17.16	.0594	34_8	
			Mean	0.0034_7	

TABLE II—Grams Bromine Per Kilo of Sea Water and the Bromine-Chlorinity Ratio Off Pribilof Islands, Southeastern Bering Sea. Lat. 57° 22′ N., Long. 170° 02′ W. and Lat. 57° 14′ N., Long. 170° 14′ W. July 16, 17, 18, 1937

Depth (*Meters*)	*Temp.* °C.	*Cl* °/₀₀	*Br* °/₀₀	*Br/Cl*
0	7.22	17.94	0.0624	0.0034_8
10	7.15	17.98	.0626	34_8
0	7.91	17.87	.0622	34_7
10	7.26	18.02	.0627	34_7
0	7.19	17.81	.0620	34_4
10	7.13	17.96	.0625	34_8
0	7.14	17.94	.0624	34_8
10	7.10	17.96	.0625	34_7
0	6.52	17.95	.0625	34_8
30	6.25	17.98	.0626	34_7
50	6.21	18.01	.0627	34_8
			Mean	0.0034_7

TABLE III—Grams Bromine Per Kilo of Sea Water and the Bromine-Chlorinity Ratio Surface Samples Northeast Pacific Beyond the Coast of British Columbia. September 24 and 25, 1938

Temp. °C.	*Cl* °/₀₀	*Br* °/₀₀	*Br/Cl*	*Location*	
12.7	17.97	0.0627	0.0034_9	Lat. 52° 06′ N,	Long. 139° 36′ W.
13.6	18.01	.0628	34_9	51° 55′ N.	138° 28′ W.
13.7	18.01	.0623	34_6	51° 52′ N.	138° 11′ W.
13.7	17.97	.0627	34_9	51° 47′ N.	138° 37′ W.
14.0	17.94	.0626	34_9	51° 44′ N.	138° 20′ W.
14.2	17.95	.0623	34_7	51° 41′ N.	138° 03′ W.
14.0	17.95	.0623	34_7	51° 38′ N.	136° 46′ W.
14.0	17.95	.0626	34_9	51° 42′ N.	136° 31′ W.
14.1	17.97	.0627	34_9	51° 32′ N.	136° 01′ W.
14.0	17.98	.0626	34_8	51° 28′ N.	135° 46′ W.
13.9	17.99	.0626	34_8	51° 24′ N.	135° 31′ W.
13.8	17.99	.0626	34_8	51° 20′ N.	135° 16′ W.
14.1	18.00	.0626	34_8	51° 16′ N.	135° 01′ W.
14.0	17.99	.0621	34_5	51° 12′ N.	134° 43′ W.
14.4	17.92	.0624	34_8	51° 08′ N.	134° 27′ W.
14.7	17.97	.0624	34_7	51° 04′ N.	134° 11′ W.
14.7	17.96	.0623	34_7	51° 00′ N.	134° 55′ W.
14.8	17.97	.0625	34_8	50° 36′ N.	133° 39′ W.
14.9	17.97	.0627	34_9	50° 52′ N.	133° 23′ W.
13.1	17.98	.0628	34_9	50° 37′ N.	132° 33′ W.
		Mean	0.0034_8		

TABLE IV—Grams Bromine Per Kilo of Sea Water and the Bromine-Chlorinity Ratio Northeast Pacific Ocean, off Coast of State of Washington. Lat. 47° 19′ N., Long. 125° 10′ W. July 8, 1939. Depth 1171 Meters

Depth (*Meters*)	*Temp.* °C.	*Cl* °/₀₀	*Br* °/₀₀	*Br/Cl*
0	15.23	17.81	0.0618	0.0034_7
10	14.66	17.83	.0617	34_8
20	11.94	17.87	.0626	35_0
30	9.03	17.98	.0627	34_8
50	8.02	18.06	.0630	34_8
75	7.70	18.29	.0631	34_8
100	7.64	18.54	.0637	34_4
200	7.58	18.78	...	..
300	5.89	18.84	.0657	34_9
400	5.54	18.87	.0654	34_8
500	5.04	18.89	.0660	34_9
600	4.58	18.92	.0661	34_9
700	4.28	18.97	.0662	34_9
800	4.00	19.00	.0662	34_8
900	3.82	19.01	.0662	34_8
			Mean	0.0034_7

TABLE V—Grams Bromine Per Kilo of Sea Water and the Bromine-Chlorinity Ratio Northeast Pacific Ocean, off Coast of State of Washington. Lat. 47° 09′ N., Long. 125° 10′ W. July 8, 1939. Depth 1427 Meters

Depth (*Meters*)	*Temp.* °C.	*Cl* °/₀₀	*Br* °/₀₀	*Br/Cl*
0	15.04	17.72	0.0617	0.0034_8
10	14.60	17.74	...	..
20	12.10	17.87	.0622	34_8
30	9.61	17.98	.0629	35_0
50	8.18	18.06	.0629	34_8
75	8.00	18.22	...	..
100	7.89	18.49	.0643	34_8
200	6.49	18.78	.0661	35_2
500	5.75	18.83	.0651	34_6
800	4.92	18.89	.0650	34_4
900	4.47	18.92	.0663	35_0
1000	4.26	18.96	.0662	34_9
1100	4.04	18.98	.0664	35_0
1200	3.76	19.01	.0664	34_9
1300	3.50	19.03	.0666	35_0
1400	3.35	19.05	.0667	35_0
			Mean	0.0034_8

TABLE VI—Grams Bromine Per Kilo of Sea Water and the Bromine-Chlorinity Ratio Strait of Juan de Fuca. Lat. 48° 182′ N., Long. 124° 03.1′ W. January 14, 1939

Depth (*Meters*)	*Temp.* °C.	*Cl* °/oo	*Br* °/oo	*Br/Cl*
0	7.94	17.28	0.0605	0.0035_0
10	7.99	17.31	.0601	34_7
20	7.93	17.34	.0607	35_0
30	7.93	17.35	.0609	35_1
50	7.83	17.55	.0611	34_8
75	7.73	17.75	.0619	34_9
100	8.21	18.24	.0636	34_9
180	8.28	18.62	.0648	34_8
			Mean	0.0034_9

TABLE VII—Grams Bromine Per Kilo of Sea Water and the Bromine-Chlorinity Ratio Waters of Puget Sound. January 13, 1939

Depth (*Meters*)	*Temp.* °C.	*Cl* °/oo	*Br* °/oo	*Br/Cl*	*Location*
0	8.26	16.48	0.0570	0.0034_6	Point No Point
10	8.47	16.60	.0573	34_8	Lat. 47° 54′ N., Long. 122° 28.7′ W.
20	8.37	16.64	.0579	34_8	
30	8.39	16.66	.0580	34_8	
50	8.36	16.69	.0581	34_8	
75	8.48	16.74	.0583	34_8	
100	..	16.77	.0582	34_7	
150	8.72	16.83	.0584	34_7	
200	8.72	16.84	.0586	34_8	
			Mean	0.0034_7	
0	8.12	16.43	0.0572	0.0034_8	Off Port Townsend
10	8.13	16.50	.0574	34_8	Lat. 48° 08′ N., Long. 122° 41.1′ W.
20	8.11	16.60	.0578	34_8	
30	8.16	16.68	.0577	34_6	
50	8.02	16.83	.0587	34_9	
75	7.99	16.85	.0588	34_9	
100	8.01	16.90	.0590	34_9	
			Mean	0.0034_8	

TABLE VIII—Grams Bromine Per Kilo of Sea Water and the Bromine-Chlorinity Ratio Various Locations in the Antarctic Ocean, Surface Waters November and December, 1930

*Cl** °/oo	*Br* °/oo	*Br/Cl*	*Location*	
19.04	0.0655	0.0034_4	Lat. 47° 20′ S.	Long. 146° 25′ E.
18.86	.0654	34_7	50° 56′ S.	147° 08′ E.
18.85	.0658	34_9	52° 25′ S.	153° 21′ E.
18.83	.0648	34_4	53° 04′ S.	154° 51′ E.
18.76	.0643	34_3	55° 55′ S.	158° 42′ E.
18.75	.0641	34_2	58° 14′ S.	160° 40′ E.
18.83	.0646	34_3	59° 16′ S.	160° 57′ E.
18.77	.0648	34_5	63° 07′ S.	166° 55′ E.
18.77	.0648	34_5	51° 06′ S.	104° 25′ E.
	Mean	0.0034_5		

* Determined by Mr. A. Howard, chemist for the British-Australian-New Zealand Antarctic Research Expedition.

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ON THE ENERGY EXCHANGE BETWEEN SEA AND ATMOSPHERE*

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FOREWORD

The present paper is a preliminary report on an investigation of the exchange of heat and water vapor between the sea and atmosphere with particular reference to the regional and seasonal variations in these quantities over the North Atlantic and North Pacific oceans. The *mean annual* evaporation over the oceans according to latitude ranges can be considered fairly well established (Wüst, 1936; Mosby, 1936; McEwen, 1938), but there has been no extensive investigation of the seasonal and intra-latitudinal variations. Since a knowledge of evaporation rates is necessary before computations of the heat exchange between sea surface and atmosphere can be attempted, as a result of lack of these data, very little is known concerning the regional variations in the total energy exchange between sea and atmosphere.

The present study approaches the problem first on the basis of energy considerations such as have successfully been employed by Mosby (loc. cit.), and mean annual evaporation values have been computed for four selected areas in the North Atlantic and North Pacific oceans. The values so derived have then been compared with the evaporation within the same areas computed on the basis of more recent theoretical considerations of the processes of interchange of water vapor within the turbulent layer near the sea surface (Sverdrup, 1937; Montgomery, 1940), when using recently published data on humidities, sea surface temperatures and wind speeds over the oceans. However, the necessary climatic data over the oceans are of doubtful

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„Mass und Zahl behalten ihre Bedeutung auch in der modernen Geography."
(N. Krebs)

accuracy, particularly with reference to humidity, and a comparison of the two methods was necessary in order to determine the statistical adjustment to be applied to these data and further, to establish the validity of the latter method of computation.

By means of the final evaporation equation obtained, seasonal values of evaporation for each five-degree square in the North Atlantic and North Pacific have been computed. In addition, through the use of the regional and seasonal evaporation values so derived, and the Bowen formula (Bowen, 1926), which allows the computation of the ratio between evaporation and heat exchange, charts have been prepared for the same areas showing the total heat and energy exchange between sea surface and atmosphere for all seasons.

I. THE AMOUNT OF HEAT ABSORBED AT THE SEA SURFACE

It is obvious that, for the oceans as a whole, the energy loss from the sea surface during the course of the year must, for all practical purposes, equal the total incoming radiation absorbed at the surface. For small areas of the sea surface it is also evident that the evaporation rate expressed in grams or centimeters of depth may be determined entirely by energy rendered available for that purpose in accordance with the equation:

Notation

E = Evaporation in grams or centimeters.
E_1 = Mean annual evaporation computed by means of the energy equation.
E_2 = Mean annual evaporation computed by means of the theoretical evaporation equation of Sverdrup (uncorrected).
Q_e = Heat of evaporation.
Q_c = Heat exchanged between sea and atmosphere through convection.
Q_a = Total energy exchange between sea and atmosphere $(Q_e + Q_c)$.
Q_t = Total solar energy reaching the sea surface.
Q_b = Total back radiation at $\bar{C}$.
Q_v = Amount of heat added to or removed from the water mass through advective processes.
L_t = Latent heat of vaporization of water $= 594.9 - 0.51t$.
Q_s = Intensity of direct sun radiation upon a surface normal to the sun's rays.
S = Solar constant (1.932 gm. cal. cm^2/min.)
Q_0 = Total incoming radiation upon a horizontal surface with clear sky.

Q_{ab} = Total heat absorbed at the sea surface.
eff Q_b = Effective back radiation with clear sky.
T = Turbidity factor (Linke).
m = Unit dry air mass at $p = 760$.
D = Ratio between direct and total radiation.
C = Percent of cloudiness.
r = Percent of incident solar heat reflected at sea surface.
R = Bowen ratio = Q_c/Q_e.
t_w = Temperature of sea surface.
t_a = Air temperature at height a.
e_w = Vapor pressure at sea surface (0.98 times e at t_w).
e_a = Vapor pressure at height a.
a = Height of observation.
H = Diffusion coefficient (0.235).
p = Pressure.
k_0 = Universal turbulence constant (von Kármán) = 0.4.
W = Friction velocity.
W_a = Velocity at height a.
z_0 = Roughness parameter (Rossby) = 0.6.
d = Thickness of laminar layer.
τ = Tangential stress of wind.
q = Specific humidity.
$\gamma_a = \dfrac{k_0}{\ln a/z_0}$
Γ = Evaporation coefficient (Montgomery).
c_p = Specific heat of air at constant pressure.
ρ = Density.
h = Solar altitude.
A = Eddy conductivity.
$\bar{k}$ = $\bar{Q}_0/\bar{h}$. (Mosby—"Richtungskoeffizient").

$$\underline{E} = \frac{Q_t - rQ_t - Q_b - Q_c - Q_v}{L_t} \qquad \text{I-(1)}$$

where Q_t is the total heat reaching the sea surface, r is the percent reflected, Q_b the heat loss through back radiation to the sky, Q_c the heat exchanged between sea and atmosphere through convective processes, Q_v the quantity of heat added to or removed from the unit water column through advective processes and L_t is the latent heat of vaporization of water at temperature t.

The intensity of the direct sun radiation upon a surface normal to the sun's rays is:

$$Q_s = S_e^{-Ta_m m} \quad \text{g. cal./cm}^2 \text{ min.} \qquad \text{I-(2)}$$

where S is the solar constant (1.932 g. cal./cm min.), e the base of natural logarithms, T the turbidity factor after the definition of Linke (1922), m the unit air mass at $p = 760$ mm. and

$$a_m = 0.128 - 0.054 \log \text{m.} \qquad \text{I-(3)}$$

It is thus only necessary to know the sun's altitude h, the turbidity factor T and, in case of necessity, the air pressure p, in order to determine the intensity of solar radiation at any point on the earth's surface. This value times the sine of $h°$ gives, then, the intensity of the direct radiation on a horizontal surface. But in dealing with the total energy received at the sea surface, Q_0, it is also necessary to take into consideration the diffuse radiation from the sky as well as the direct radiation from the sun. From observations on the *"Maud"* expedition, Mosby (loc. cit.) obtains the following relationship between direct and total radiation:

$$D = 0.88 - \frac{2.4}{h} \qquad \text{I-(4)}$$

and for approximate values of T, obtains:

$$Q_0 = \frac{1}{D} Q_s \cdot \sin h \qquad \text{I-(5)}$$

It has been shown by Mosby, however, that the total incoming radiation on a horizontal surface at sea level can be related to the average solar altitude by the very convenient linear equation:

$$Q_0 = k\bar{h} \text{ g. cal./cm}^2 \text{ min.} \qquad \text{I-(6)}$$

The values of k vary somewhat with latitude within a range of 0.023 to 0.027 due to the varying water vapor in the atmosphere. The product $\overline{kh}$, then, gives the yearly average of the total incoming radiation with a clear sky in the stated units.

Ångstrom (1922), Kimball (1928) and others have determined an approximate relation between daily totals of solar radiation and the average cloudiness, the values for the ratio between radiation with clear sky and totally cloud-covered sky falling within the range 0.22 to 0.30. The differences are largely determined by the nature and thickness of the cloud cover and, perhaps, also by the percentage of cloudiness and solar altitude. Kimball presents 0.29 as the most likely average value. Using this ratio throughout he finds:

$$Q_t = Q_0 \, [0.29 + 0.71 \, (1 - \bar{C}/100)] \qquad \text{I-(7)}$$

where C is the percentage of sky covered by clouds. This allows the determination of the average incoming radiation for any latitude by means of the following formula:

$$Q_t = kh \, [1 - 0.71 \, (\bar{C}/100)] \qquad \text{I-(8)}$$

This formula, however, determines only that amount of heat incident upon the sea surface; a reduction is necessary to allow for the amount reflected back to the sky. According to Schmidt (1915), the percent reflected, r, varies according to latitude from a minimum of 3.3 at the equator to 8.0 at the poles. These data are given in tabular form by this author and also by Ångstrom (1920) and Mosby (1936). From this it follows that the total solar heat absorbed at the sea surface is:

$$Q_{ab} = Q_t \, (1 - r) \qquad \text{I-(9)}$$

Now, of the total heat absorbed by the water mass, some will be radiated back to space, the amount of which depends upon the absolute temperature of the water surface, the moisture content of the atmosphere and upon the nature and extent of cloud cover. Ångstrom (loc. cit.) has computed values for the back radiation at various temperatures and humidities and the data in the present calculations have been obtained from that source. As suggested, however, the values for the effective back radiation ($eff \, Q_b$) so obtained must be reduced to allow for the screening effect of any cloud cover present. According to Asklöff (1920), this correction factor amounts to $(1 - 0.0083 \, \bar{C})$. The final value must be further reduced (Ångstrom) by 6% to allow for internal reflection. Thus, the average back radiation is represented by the following formula:

$$Q_b = 0.94 \, [eff \, Q_b \, (1 - 0.0083 \, \bar{C})] \qquad \text{I-(10)}$$

The difference $Q_{ab} - Q_b$ represents, therefore, the amount of heat available in the sea for evaporation and for heating the air. It is necessary, as the first step in the computations, to disregard the advection factor. In order to determine the amount of heat used up in the evaporation process it is now only necessary to determine the proportion of energy given off as sensible heat to the atmosphere. All of the investigators of the problem referred to here have assumed a constant value for this proportion but it will be shown that the ratio Q_c/Q_e is highly variable, both with respect to season and with respect to the regional distribution.

II. THE EXCHANGE OF HEAT BETWEEN SEA AND ATMOSPHERE

Mosby (loc. cit.), concerning the exchange of sensible heat between sea surface and atmosphere, states:

> „Um die mittlere Verdunstung bestimmen zu können, müssen wir nunmehr also noch diejenige Wärmemenge kennen, die durch turbulente Konvektion von der Meeresoberfläche zur Atmosphäre verfrachtet wird. Diese Wärmemenge hangt von vielen Verhältnissen ab, in erster Linie von dem Temperaturunterschied Wasser—Luft und von der Windstärke; es ist jedoch heute kaum möglich, sie gehau zu bestimmen."

In his determinations of evaporation from the oceans, Mosby assumes an arbitrary, but reasonable, value of 10% as the proportion of available energy given off as sensible heat to the atmosphere. He appears not to have been acquainted with the work of Bowen (1926), who has established the relation between evaporation and heat exchange at a water surface.

In considering the rates at which heat and water vapor are transported across the lower and upper surfaces of a volume of air above a body of water, Bowen has determined that the ratio, R, of the heat loss by conduction to that by evaporation can be obtained by means of the following formula:

$$R = 0.50 \left(\frac{t_w - t_a}{e_w - e_a} \right) \frac{p}{760} \qquad \text{II-(1)}$$

where t_w and t_a are the temperatures in $°C$ of the water and air, respectively, e_w and e_a are the vapor pressures in millimeters of the water surface and air and p is pressure (mm.). Bowen's derivation of the above formula is too lengthy to be reproduced here but the same equation can be derived quite simply as follows (Sverdrup et al, 1941):

Since the amount of water vapor, F, which is transported upward through the unit surface per second is

$$F = -A \frac{0.623}{p} \frac{de}{dz} \qquad \text{II-(2)}$$

where A is the eddy conductivity, then the amount of latent heat transported as water vapor is

$$Q_e = -L_t A \frac{0.623}{p} \frac{de}{dz} \qquad \text{II-(3)}$$

The amount of sensible heat transported away from the sea surface is

$$Q_c = -c_p A \frac{dt_a}{dz} \qquad \text{II-(4)}$$

where c_p is the specific heat of air at constant pressure. Therefore:

$$R = \frac{Q_c}{Q_e} = \frac{c_p}{L_t} \frac{p}{0.623} \frac{dt_a}{dz} \Big/ \frac{de}{dz} \qquad \text{II-(5)}$$

Setting $c_p = 0.237$ and $L_t = 585$, one obtains

$$R = 0.49 \frac{p}{760} \frac{dt_a}{dz} \Big/ \frac{de}{dz} \qquad \text{II-(6)}$$

Replacing the gradients dt_a/dz and de/dz by $(t_w - t_a)$ and $(e_w - e_a)$, one finds

$$R = 0.49 \frac{p}{760} \left(\frac{t_w - t_a}{e_w - e_a} \right) \qquad \text{II-(7)}$$

Since the present calculations require sea level pressures throughout, obviously the pressure term can be neglected, thus:

$$Q_c = RQ_e = \frac{RQ_a}{1+R} \qquad \text{II-(8)}$$

$$Q_a - Q_c = \frac{Q_a}{1+R} = EL_t \qquad \text{II-(9)}$$

and from (7) and (9)

$$E = \left[\frac{Q_a}{1 + 0.49 \left(\frac{t_w - t_a}{e_w - e_a} \right)} \right] \Big/ L_t \qquad \text{II-(10)}$$

III. COMPUTATIONS OF THE MEAN ANNUAL EVAPORATION FOR SELECTED AREAS IN THE NORTH ATLANTIC AND NORTH PACIFIC OCEANS BY MEANS OF ENERGY CONSIDERATIONS

In order to compare the values for evaporation obtained through the use of the energy equations given above with those determined by means of the more recent theoretical considerations of the interchange of water vapor and heat within the turbulent layer of the atmosphere, computations of the mean annual evaporation have been made for four selected areas in the North Atlantic and North Pacific

oceans. Since, as a first step, it is necessary to disregard the advection factor, Q_v, the areas selected were those within which the dia-latitudinal transport of ocean waters is at a minimum. For this reason, coastal areas have been eliminated from consideration. The two latitude ranges chosen were between 40° N. and 45° N. and between 20° N. and 25° N. In the North Atlantic, between 40° N. and 45° N., the evaporation was computed for the area bounded by meridians 20° W. and 50° W. and in the latitude range 20° N.–25° N., by the meridians 25° W. and 65° W. In the North Pacific, between 40° N. and 45° N., the computations have been made for the area bounded by meridians 140° W. and 160° E. and between latitudes 20° N.–25° N., by 130° W. and 140° E.

For the marine climatic data entering into the computations, the writer is indebted to the New Orleans and Washington, D. C., offices of the U. S. Weather Bureau for the loan of manuscript charts presenting recently compiled data on sea surface and air temperature, wet-bulb depression, wind speed and cloudiness over the oceans. Abridged forms of some of these charts appear in the *Atlas of Climatic Charts of the Oceans* (U. S. Weather Bureau, 1938).

The results of the computations are given in Table I. The average annual evaporation in the North Atlantic between latitudes 40° N. and 45° N. proves to be 84 cm/yr.; in the North Pacific, 68 cm/yr. Within the latitude range 20° N.–25° N., the evaporation is 168 cm/yr. in the Atlantic Ocean and 148 cm/yr. in the Pacific Ocean. These results would indicate that evaporation is greater at all latitudes in the North Atlantic than in the North Pacific. The value for the latitude range 40° N.–45° N. in the Atlantic is 16 cm/yr. greater than for the similar latitude range in the Pacific while it is 20 cm/yr. greater for the latitude range 20° N.–25° N. Analysis of the data in Table I indicates that the higher evaporation in the North Atlantic is largely due to the smaller amount of heat lost to the atmosphere rather than to any significant excess of energy received. The mean annual cloudiness within the latitude range 40° N.–45° N. is, however, somewhat higher in the North Pacific than in the North Atlantic which would account for a small proportion of the difference in evaporation at these higher latitudes.

TABLE I—Computations of Evaporation for

	Lat.	*Long.*	k	$\bar{h}°$	$\bar{C}$ (%)	Q_t/Q_o	Qt	r (%)	Q_{ab}
Atlantic	40°N–45°N	25°W–10°W	0.024	14.7	64	0.546	0.193	4.8	0.184
Ocean	20°N–25°N	25°W–65°W	0.023	18.2	46	0.673	0.282	3.8	0.271
Pacific	40°N–45°N	140°W–160°E	0.024	14.7	70	0.506	0.178	4.8	0.169
Ocean	20°N–25°N	130°W–140°E	0.023	18.2	46	0.675	0.283	3.8	0.272

IV. COMPUTATIONS OF THE ANNUAL EVAPORATION BY MEANS OF AVAILABLE DATA ON WIND SPEEDS, HUMIDITIES AND TEMPERATURES OVER THE OCEANS

As an outgrowth of the researches on turbulent mixing in the lower layers of the atmosphere, particularly by Taylor (1932), Prandtl (1935), von Kármán (1935), Rossby (1932; 1935) and Sverdrup (1936), it has been possible to prepare formulae showing evaporation as a function of moisture concentration and wind movement within the turbulent layer. Such formulae have been constructed by Sverdrup (1937) and Montgomery (1940) but since their discussions are presented in readily available publications, no attempt will be made in this paper to enter into the derivations of the equations.

However, the basic assumptions are that the transport of water vapor between sea surface and atmosphere takes place by ordinary diffusion through a shallow boundary layer and by eddy conductivity above this layer. It is also shown that for the unstable atmosphere, i. e., where the potential temperature decreases with height, the eddy conductivity, itself, is a linear function of height and of the roughness character of the sea surface. It is also shown that, under these conditions, the vapor pressure is a function of the logarithm of height. According to Sverdrup:

$$E_2 = \frac{H \dfrac{0.623}{p} \rho \ (e_w - e_a) W}{\dfrac{H}{K_0} \ln \dfrac{a + Z_0}{d + Z_0} + dw} \qquad \text{IV-(1)}$$

where H is the diffusion coefficient, p is the atmospheric pressure, ρ is density, a altitude of observation, k_0 is von Kármán's constant (0.4), W is the friction velocity where

$$W = \sqrt{\frac{\tau}{\rho}} = \frac{k_0}{\ln \dfrac{a + z_0}{z_0}} \quad W_a = \frac{0.165}{\log \dfrac{a \pm z_0}{z_0}} \quad W_a$$

Selected Areas by Means of the Energy Equations

$\bar{t}w$ (°C)	$\bar{t}a$ (°C)	$\bar{e}a$ (mm)	$\bar{e}w$ (mm)	$eff Q_b$	Q_b	$Q_{ab} - Q_b$	R	Q_c	Q_e	$\bar{L}_t$	E_1 (cm/yr.)	
16.4	16.1	11.9	13.7	0.187	0.083	0.101	0.076	0.007	0.094	586.9	84.0	*Atlantic*
24.8	25.2	19.8	23.0	0.166	0.096	0.175	−0.056	−0.010	0.186	582.3	167.5	*Ocean*
11.1	10.4	8.5	9.8	0.186	0.074	0.095	0.249	0.019	0.076	589.8	67.7	*Pacific*
24.6	23.9	18.6	22.8	0.166	0.097	0.176	0.073	0.012	0.164	582.4	148.2	*Ocean*

W_a is the velocity at level a and z_0 is the roughness parameter (0.6,—Rosby) and

$$dW = \frac{dW_a\ k_0}{\ln \frac{a}{z_0}} = 4.12$$

According to Montgomery:

$$E_2 = \rho k_0\ \gamma_a\ \Gamma\ (q_w - q_a)\ W_a \qquad \text{IV-(2)}$$

where $\gamma_a = \dfrac{k_0}{\ln \dfrac{a}{z_0}}$ and Γ is his "evaporation coefficient." Both γ_a and Γ depend upon the hydrodynamic character of the sea surface but the dependence is such that the product $\gamma_a\ \Gamma$ is very nearly constant for either rough or smooth surface. If this is true, equation IV-(2) can be reduced to a formula of the type:

$$E_2 = K\ (e_w - e_a)\ W_a \qquad \text{IV-(3)}$$

Equation IV-(1) also reduces to a formula of this type.

Since $d = \dfrac{\ln \dfrac{a}{z_0}\, dW}{k_0\, W_a} = \dfrac{71.2}{W_a}$

when $z_0 = 0.6$ and $a = 600$

Then, at $W_a = 500$; $\quad d = 0.14$

$W_a = 1500$; $\quad d = 0.05$

Taking $H = 0.235$ and $p = 1000$, from equation IV-(1):

$$E_2 = \frac{0.235 \cdot 0.623 \cdot 1.2 \times 10^{-6} \cdot \dfrac{0.4}{6.9}\ (e_w - e_a)\ W_a}{0.6 \cdot 6.6 + 4.1}$$

$$= 1.25 \times 10^{-9}\ (e_w - e_a)\ W_a \qquad \text{IV-(4)}$$

where e is in millibars and W_a is in meters per second at a height of six meters.

The 24-hour evaporation in millimeters or decigrams is therefore given by:

$$E_2 = 0.108\ (e_w - e_a)\ W_a \qquad \text{IV-(5)}$$

Equation IV-(1) applies to the adiabatic atmosphere but under conditions of atmospheric stability, the computed values for E_2 will be somewhat too great. However, when considering evaporation over periods of a day or longer, no serious error is introduced as the amounts computed under stable conditions are small when compared with those computed under adiabatic conditions.

Through the use of equation IV-(5), computations of the mean annual evaporation have been made for the same four areas in the North Atlantic and North Pacific. The results are given in Table II. As

TABLE II—Mean Annual Evaporation for Selected Areas Computed by Means of Equation (5), E_2, Compared with the Mean Annual Evaporation, E_1, Computed by Means of the Energy Equations

	Lat.	*Long.*	E_1 (*cm/yr.*)	E_2 (*cm/yr.*)	E_1/E_2	E ($E_2 \times 1.32$) (*cm/yr.*)
North Atlantic	40°N–45°N	20°W–50°W	84.0	88.3	0.95	116.6
North Atlantic	20°N–25°N	25°W–65°W	167.5	94.2	1.78	124.3
North Pacific	40°N–45°N	140°W–160°E	67.7	55.8	1.21	73.7
North Pacific	20°N–25°N	130°W–140°E	148.2	113.2	1.31	149.4

was expected, the values derived by the latter method are lower than those computed by means of the energy equations. The mean annual evaporation for the four areas proves to be 87.9 cm/yr. when computed by equation IV-(5) and 116.8 cm/yr. when computed by means of the energy equations. Most of this difference can no doubt be assigned to the inaccuracies of the humidity measurements at sea for it would be expected that the mean of a large number of such observations would be too high, the tendency of error in the individual observations being in this direction. This would result, of course, in the computed values of evaporation being too low. However, if these differences resulting from the inadequacy of the observational data are consistent, it is possible to prepare a correction factor to be applied to equation IV-(5) such that the two methods can be brought into agreement. The ratios E_1/E_2 given in the sixth column of Table II, indicate that this factor is of the order of 1.32. Thus equation IV-(5) becomes:

$$E = 0.143\ (e_w - e_a)\ W_a \qquad \text{IV-(6)}$$

It now becomes of interest to examine the nature and causes of the remaining differences $(E_1 - E)$.

In the use of the energy equations under section III for computing E_1, it was necessary to disregard the advection term, Q_v, except to the extent that the sea surface temperature data entered into the computations of Q_b and R. For this reason, it would be expected that the evaporation computed by such methods would be too low at high latitudes and too high at low latitudes. An examination of the data in the last column of Table II shows this to be particularly true in the case of the North Atlantic where the dia-latitudinal transport of surface waters is greater than in the North Pacific. Thus in the North Atlantic, on account of the great northerly transport in the west portion, the greater heat transport is from south to north. In the case of the North Pacific, on the other hand, the northerly transport in the west portion is more nearly balanced by the southerly transport in the east portion. If we set

$$Q_v \cong \Delta Q_{1,2} = (E\ L_t + Q_c) - (E_1 L_t + Q_{c1})$$

then for the Atlantic between latitudes 40° N–45° N., Q_v becomes 20533 g. cal. cm^2/yr. and for the latitude range, 20° N.–25° N., — 23811 g. cal. cm^2/yr. In the Pacific, on the other hand, Q_v is only 444; g. cal. cm^2/yr. for the latitude range 40° N.–45° N. and still shows a slight positive net heat transport between latitudes 20° N.–25° N. of 626 g. cal. cm^2/yr. The ratios of Q_v for North Atlantic and North Pacific indicate that the heat transport into the latitude range 40° N.–45° N. of the Pacific is only 21% as great as the similar transport in the Atlantic within the same latitude range, while in the latitude range 20° N.–25° N. in the Pacific, the inflow of heat is very nearly balanced by the outflow. These results are in line with present information concerning the transport of surface waters in both oceans. It should perhaps be repeated, that since the areas were so chosen as to exclude the coastal sections, the net heat transport for the entire latiude ranges represented would differ from the figures given above.

V. COMPUTATIONS OF THE SEASONAL EVAPORATION, HEAT EXCHANGE AND TOTAL ENERGY EXCHANGE FOR FIVE-DEGREE SQUARES IN THE NORTH PACIFIC AND NORTH ATLANTIC

Through the use of equation IV-(6), it is now possible to compute the mean seasonal evaporation for five-degree squares in the North Atlantic and North Pacific by means of the previously-mentioned data on sea surface and air temperatures, wet-bulb depressions and wind speeds over the oceans. The values for e_w have been obtained

by multiplying the vapor pressures corresponding to the surface temperature t_w, by 0.98 in order to take the effect of salinity into account. The values for e_a have been obtained from the Smithsonian Tables which present vapor pressure as a function of the difference between wet- and dry-bulb temperatures.

Then from the evaporation quantities so determined, and the values of $t_w - t_a$ and $e_w - e_a$, the seasonal and annual values for Q_c and Q_a have been computed for the same areas.

Since the climatic data at hand presented temperatures and wet-bulb depressions in degrees Fahrenheit and the wind speeds in knots, it was convenient to adjust the several equations to the same units such that:

$$E = 0.249\ (e_w - e_a)\ w_a \qquad \text{gms. cm}^2\text{/day} \qquad \text{V-(1)}$$

$$Q_e = 145.4\ (e_w - e_a)\ w_a \qquad \text{gm. cal. cm}^2\text{/day}\ (L = 585) \qquad \text{V-(2)}$$

$$Q_c = 0.01 \left(\frac{t_w - t_a}{e_w - e_a} \right) Q_e \qquad \text{gm. cal. cm}^2\text{/day} \qquad \text{V-(3)}$$

$$Q_a = Q_e + Q_c \qquad \text{gm. cal. cm}^2\text{/day} \qquad \text{V-(4)}$$

in which e is in inches-Hg, t is in degrees Fahrenheit and w_a is in standard knots.

The charts prepared from the mean seasonal and annual values of E and Q_c computed for five-degree squares in the North Pacific and North Atlantic are primarily of interest in the meteorological field which is not only concerned with the amount of heat added to or subtracted from the lower layers of the atmosphere, but also of the amount of latent energy in the form of water vapor which is rendered available for the various meteorological processes. On the other hand, the charts showing the seasonal and regional values of Q_a are of particular interest to the field of oceanography since the loss or gain of energy from the sea surface through either evaporation or conduction is entirely in the form of heat. The charts for E are of interest in this field only to the extent that the evaporation, through its effects upon the surface salinity, serves to alter the distribution of density within the oceans. However, it is not practicable to include the 15 charts necessary to show the seasonal and annual values of E, Q_c and Q_a in the present short paper. It is expected that these will be presented in a later, and more complete, report.

An examination of the charts for E shows that the evaporation in both oceans is nearly everywhere greatest in winter and least in summer; that the regions of greatest evaporation are on the western

sides of the oceans and about the southern margins of the North Atlantic and North Pacific semi-permanent high pressure fields. Evaporation on the eastern sides of the oceans is generally smaller. The absolute maximum evaporation for all oceans occurs in the North Atlantic in winter within the Gulf Stream between latitudes 35° N.–40° N. and about 700 miles east of the Virginia Capes, where the average evaporation is about 1.14 gms. cm^2/day. The maximum value in the North Pacific, 0.94 gms., on the other hand, occurs farther south and nearer the coast within the Kuroshio between latitudes 25° N.–30° N. and approximately 300 miles northeast of Formosa. The evaporation in winter along the eastern sides of both oceans is relatively much smaller, for example, in the North Pacific it is only 0.15 gms. cm^2/day off the coast of British Columbia and Central California and 0.14 gms. off the Southern California Coast. Although the values increase slightly southward, the evaporation is still only 0.18 gms. in the Gulf of Panama. In the North Atlantic off the Irish Coast, the winter value is 0.30 gms. and off the northern coast of Africa, 0.36 gms. The average winter evaporation in the Gulf of Mexico and the Caribbean Sea is of the order of 0.40 gms.

In spring the evaporation decreases considerably along the western sides of the oceans, the influence of the Kuroshio upon evaporation in the North Pacific having almost disappeared while the similar effect of the Gulf Stream in the North Atlantic has become considerably less. The evaporation in the south-central portion of the tropical North Pacific, however, increases slightly during this season and reaches its maximum value. The evaporation along the equator in the eastern North Atlantic also increases slightly but at the same time decreases to the minimum seasonal value within the corresponding region in the North Pacific. In the higher latitudes, the central and western portions of the equatorial regions and along the eastern sides of the oceans, the decrease in evaporation is small.

In summer the evaporation reaches its lowest value over most of the North Pacific and North Atlantic. The values, however, increase slightly during this season within the eastern equatorial regions of both oceans and reach their maximum at this time. The decrease in evaporation within the tropical belts of high evaporation is small. The quantities, nevertheless, remain positive for all areas except in the North Pacific north of latitude 55° and west of 160° W., and in the North Atlantic waters immediately surrounding Labrador, where a slight net condensation is computed for this season. The effect of the Kuroshio in the western North Pacific has now completely disappeared and the computed values of evaporation for the eastern

and western sides of this ocean are of the same order of magnitude. The effects of the Gulf Stream upon evaporation are, however, still in evidence and the evaporation on the west side of the North Atlantic remains greater than in the eastern portions.

In the autumn the evaporation nearly everywhere increases and is greater than during either spring or summer except in the south-central tropical regions in both oceans where the minimum values for these respective areas occur. Evaporation also decreases from its summer maximum in the eastern equatorial regions of both oceans. The effect of the Kuroshio upon evaporation in the North Pacific again becomes evident and evaporation within the Gulf Stream of the North Atlantic now reaches values as high as 0.88 gms. off the Virginia Capes.

Thus, in general, it is shown that the regions of greatest evaporation are those within which the northerly transport of surface waters is greatest, i. e., along the western sides of the oceans within the Kuroshio and Gulf Stream. Conversely, the regions of least evaporation are those of southerly flow or drift, namely, along the eastern sides of the oceans and the extreme northwestern coastal areas. The only exceptions to this general statement appear within the south-central tropical areas of both oceans where, in the northern part of the trade wind region, a relatively high evaporation is associated with the dry descending air currents accompanying the semi-permanent fields of high pressure, and in the eastern equatorial regions, where relatively low values are associated with the northerly flowing ocean currents which in these regions cross the equator resulting in an influx of cold water from the higher latitudes in the Southern Hemisphere.

In the North Atlantic Ocean, the center of the tropical or trade-wind area of maximum evaporation remains nearly stationary during all seasons except autumn, when it quite largely disappears, with its mean position located between latitudes 10° N.–20° N. and approximately one-third the distance between the coasts of South America and Africa. In the North Pacific Ocean, on the other hand, this tropical area of high evaporation is considerably better developed and its center varies in position with the seasons. In winter the maximum evaporation within this area is 0.70 gms. cm^2/day and the center is located between 5° N.–10° N. and 170° W.–175° W. In spring the maximum increases to 0.72 gms. and the center is displaced northeastward to the area between 10° N.–15° N. and 150° W.–155° W. In summer the maximum evaporation in this region decreases to 0.62 gms. and the center is further displaced northeastward to the area between 15° N.–20° N. and 145° W.–150° W. In the autumn, how-

ever, the maximum evaporation within this area falls to its lowest value (0.56 gms.) and the now less well-defined center is displaced southward to the general region on the equator between latitudes 0° N.–5° N. and between 135° W.–150° W. This clockwise migration of the tropical center of high evaporation in the North Pacific appears to correspond to the similar movement of the center of the North Pacific high pressure field.

The average annual values of evaporation for the various latitude ranges are given in Figure 18 and are compared with similar values presented by other investigators. The computed annual values appear to be in excellent agreement with those given by Wüst (1936); the ratio $E/E_{\text{Wüst}}$ for all latitudes being 1.00 for the North Atlantic, 1.06 for the North Pacific and 1.03 for both oceans. The agreement with McEwen's (1938) values is also good but the values are lower than those given by Cherubim (1931). The agreement with Mosby's (1936) values is good for the latitude ranges north of 25°, but his values are considerably higher for the lower latitudes. The greatest departures between the present values and those previously determined by other investigators of the problem, appear within the latitude range 45° N.–50° N., where the previous values given have been somewhat higher, and within the latitude range 35° N.–40° N., where the previous values assigned have been lower. In general, these curves show that evaporation in the North Atlantic is at a maximum within the latitude range 35° N.–40° N. with a well defined secondary maximum between latitudes 15° N.–20° N. In the North Pacific, the maximum annual evaporation occurs between latitudes 20° N.–25° N. with evidence of a slight secondary maximum between latitudes 35° N.–40° N. With regard to the curves shown, it is perhaps well to remark that too much dependence should not be placed upon the values north of latitude 55° since the areas involved in the computations for these regions are rather small.

The isolines for Q_c over the oceans show, roughly, the same configuration as shown for E, these values being at their maximum in winter and along the western sides of the oceans, with the major exception that there exist no tropical areas of maximum heat exchange to coincide with the areas of maximum evaporation which appear associated with the North Pacific and Azores high pressure systems.

The absolute maximum value for Q_c (260 g. cal. cm^2/day) occurs in winter in the North Atlantic in about the same location as the center of maximum evaporation. In the North Pacific, the maximum winter value (240 g. cal. cm^2/day) occurs between 35° N.–40° N. and 150° E.–155° E. with a secondary area of maximum heat exchange northeast

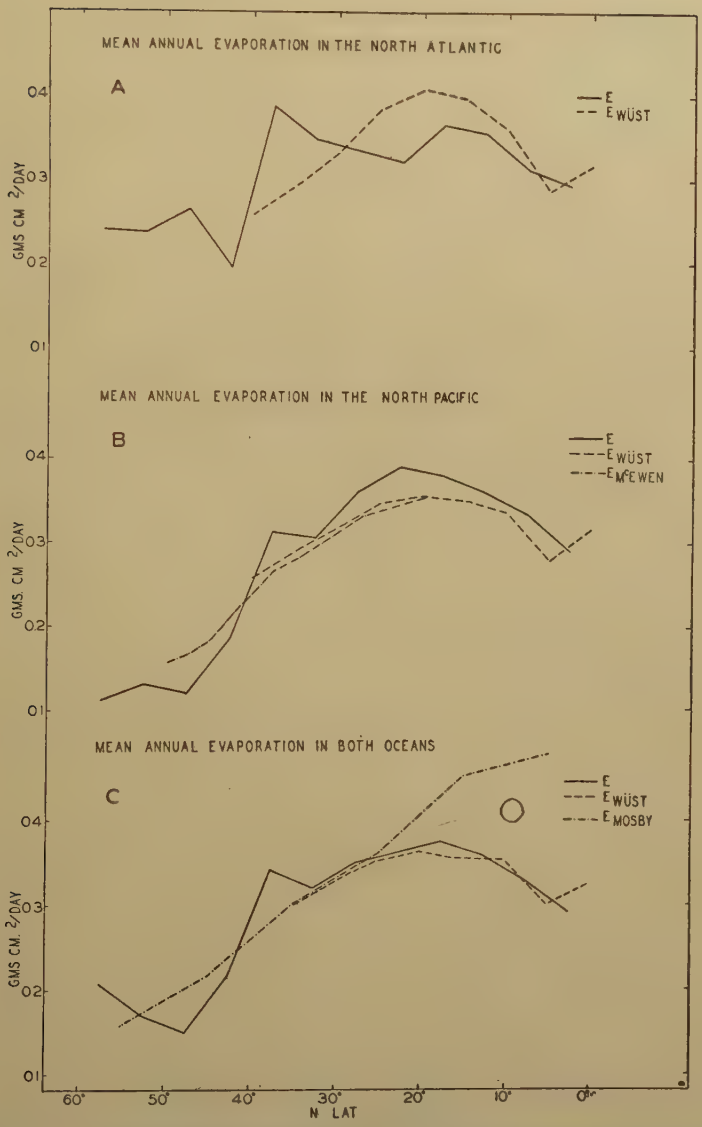

Figure 18. A. The mean annual evaporation in the North Atlantic compared with the values previously determined by Wüst. B. The mean annual evaporation in the North Pacific compared with the values previously determined by Wüst and McEwen. C. The mean annual evaporation in both oceans compared with the values previously determined by Wüst and Mosby for all oceans.

of Formosa coinciding with the area of maximum evaporation. The North Pacific values for Q_c during winter are positive for all regions except for several small areas near the equator. In the North Atlantic, however, the charts show that even during the winter, Q_c is negative for approximately one third the area of that ocean, i. e., roughly southeast of a line drawn northeast from Trinidad, B. W. I., to lat. 40° N., long. 20° W. and thence eastward.

In spring Q_c decreases generally in all areas but still remains positive for most of the North Pacific. Values as high as 116 g. cal. cm^2/day occur within the Gulf Stream but Q_c is negative for all that area of the North Atlantic roughly southeast of a line drawn from Panama to lat. 60° N., long. 10° W. A value of -57 g. cal. cm^2/day is computed for the area immediately off the coast of Africa between 15° N. and 20° N. In summer the values for Q_c become insignificant nearly everywhere in the North Pacific except in the southeastern equatorial regions where values as high as 40 g. cal. cm^2/day are determined, and are negative everywhere north of latitude 40°, off the California Coast and, surprisingly, also off the Asiatic coasts south to the Philippine Islands. In the North Atlantic in summer, Q_c is negative everywhere except within a narrow zone along the western side extending northeastward from Florida to lat. 40° N., and within the Gulf of Mexico and Caribbean Sea.

In autumn, the values for Q_c increase considerably in both oceans and the distribution is very nearly the same as in spring except that the negative values in the eastern portions of the North Atlantic are not as great.

The mean annual values of Q_c arranged by latitudes are shown in Figure 19A. These curves indicate that the heat exchange between sea and atmosphere per unit area is greater in the North Pacific than in the North Atlantic, particularly south of the latitude range 35° N.–40° N. From minima within the latitude range 45° N.–50° N., Q_c in both oceans increases to the latitude range 35° N.–40° N. and then decreases toward the equator, becoming negative in the North Atlantic south of lat. 25°.

The charts showing the total energy exchange (Q_a) between sea surface and atmosphere are, of course, composite charts of the values of Q_e and Q_c, therefore, in general configuration they are quite similar to the charts for E, except that the west-east gradient in the energy exchange is intensified due to the high values for Q_c over the western sides of the oceans and the low values over the eastern sides. Conspicuous areas of maximum energy exchange over the south-central tropical areas of both oceans coincide with the tropical areas of high

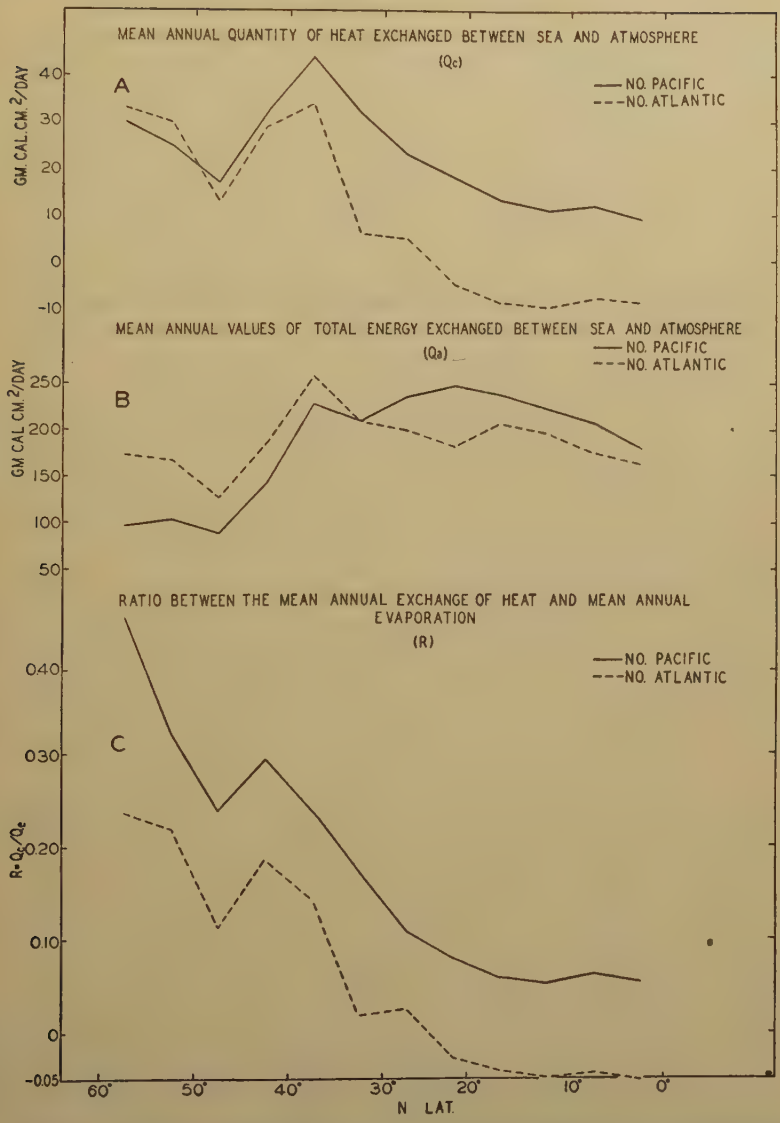

Figure 19. A. Mean annual values for Q_c in the North Atlantic and North Pacific oceans. B. mean annual values for the total energy exchange (Q_a) in the North Atlantic and North Pacific oceans. C. Ratio (R) between the mean annual values for Q_c and Q_e in the North Atlantic and North Pacific oceans.

evaporation as would be expected. In the North Atlantic in winter, Q_a reaches values as high as 937 g. cal. cm²/day within the Gulf Stream off the Virginia Capes. In the North Pacific, the maximum energy exchange occurs in winter within the Kuroshio northeast of Formosa, coinciding with the area of maximum evaporation, where a value of 728 g. cal. cm²/day is computed, with a secondary area of maximum energy exchange (718 g. cal. cm²/day) between latitudes 35° N. and 40° N. approximately 350 miles east of Japan.

The annual values for Q_a arranged by latitudes are given in Figure 19B, which show that the maximum annual energy exchange in the North Atlantic takes place between latitudes 35° N. and 40° N., and in the North Pacific, between latitudes 20° N. and 25° N. with a slight secondary maximum between latitudes 35° N. and 40° N.

Sverdrup (1940) has investigated (by means of the energy equation) the annual variation of the loss of energy from the sea surface in the North Atlantic and North Pacific within two small areas where the necessary observational data were available. In the North Atlantic off the Bay of Biscay at 47° N. and 12° W., he found the energy loss to be greatest in late fall or early winter, becoming zero during summer and with a secondary minimum in February and a secondary maximum in March. These conclusions are in line with the present results for the same region with the exceptions, of course, that neither the secondary minimum nor the secondary maximum could be expected to show in the present calculations. He found the average annual loss of energy in this area to be of the order of 140 g. cal. cm²/day, which is somewhat greater than the present 90 g. cal. average computed for the five degree square immediately west of Sverdrup's point of computation. However, it is to be noted that Sverdrup's calculations have been made for a very small area and one which is quite near the coast, therefore it would be expected that his values would depart somewhat from those computed for a larger area farther westward.

In the North Pacific, Sverdrup has determined, by similar method, the annual variation in the energy loss from the sea surface for a small area within the Kuroshio along the meridian 135° E. between latitudes 31° 50′ N. and 33° 30′ N. He found the energy exchange here also to be at its maximum in winter, the average during December, January and February being approximately 1100 g. cal. cm²/day. This value is considerably higher than the present calculations give for the same general area, i. e., about 600 g. cal. However, again it must be considered that Sverdrup's computations have been made only for a very small area within which the temperature gradients were found to be very steep and should, therefore, give greater values than those computed for the five degree square.

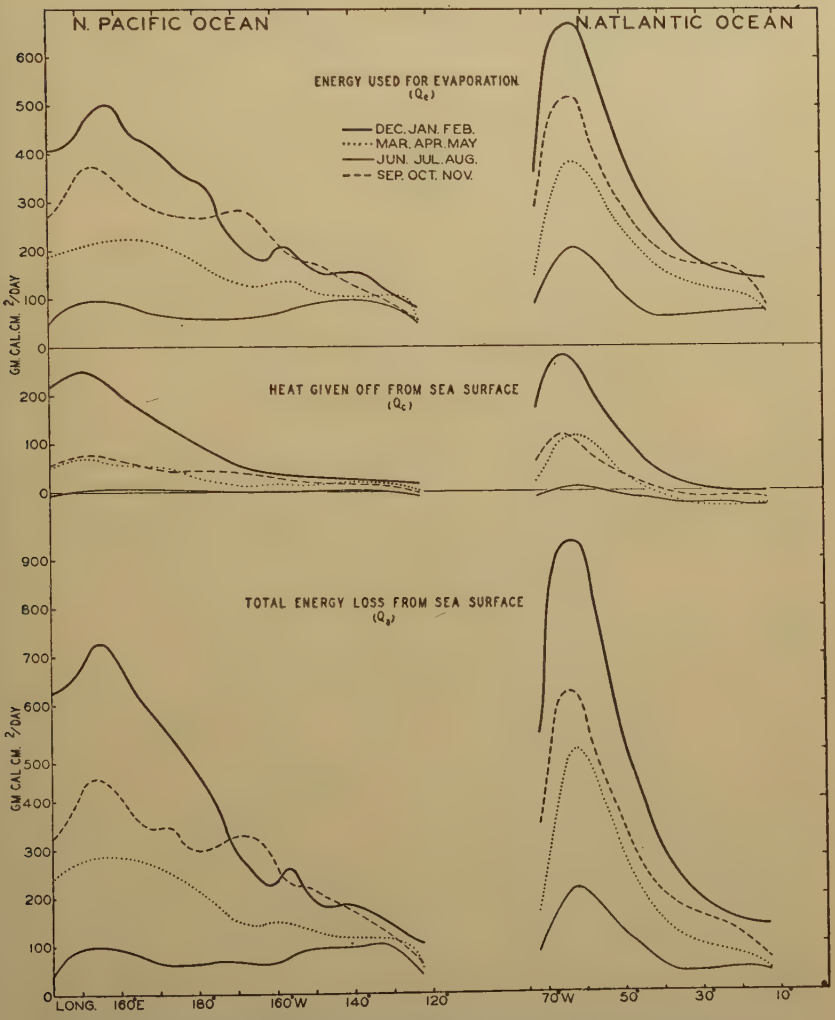

Figure 20. Seasonal variations in evaporation, heat exchange and the total energy exchange within the latitude range 35° N.–40° N. in the North Atlantic and North Pacific oceans.

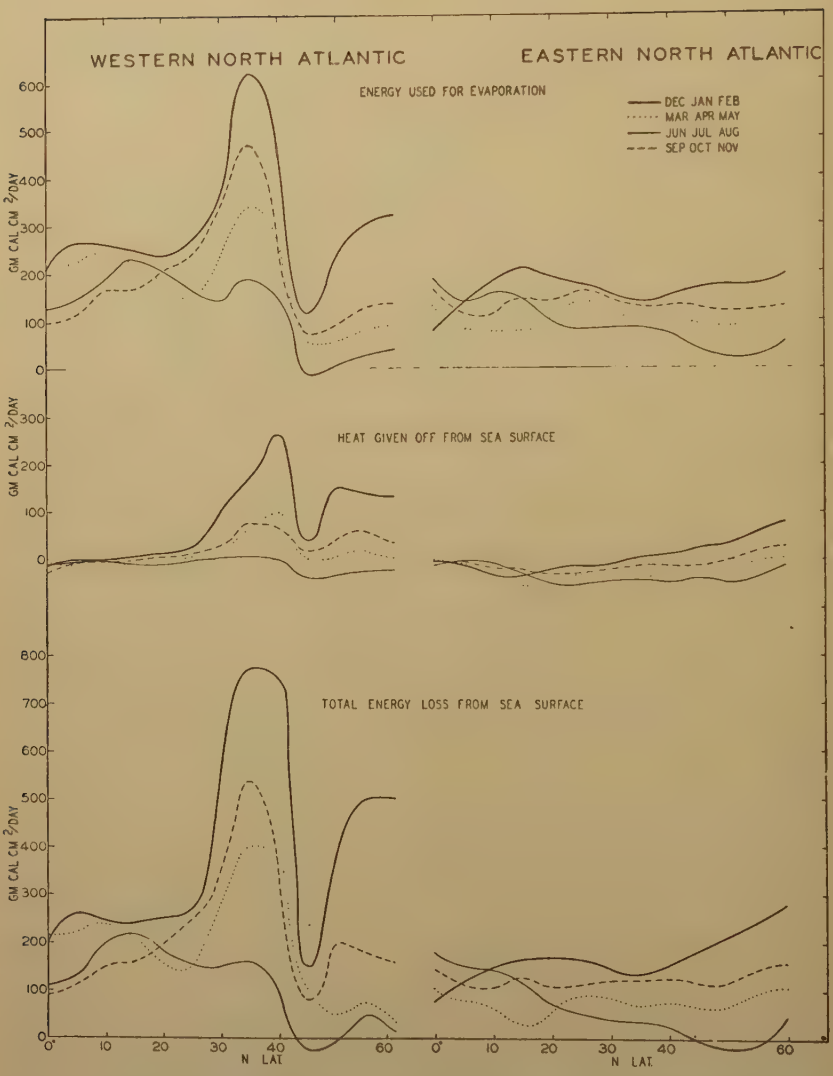

Figure 21. Seasonal variations in evaporation, heat exchange and the total energy exchange according to latitude along the western and eastern sides of the North Atlantic ocean.

The seasonal variations in the quantities Q_e, Q_c and Q_a across the North Pacific and North Atlantic within the latitude range 35° N.–40° N. are shown in Figure 20. These curves serve to illustrate the large seasonal variations in these quantities over the western portions of both oceans, i. e., within the Kuroshio and Gulf Stream, and the relatively much smaller seasonal variations over the eastern portions.

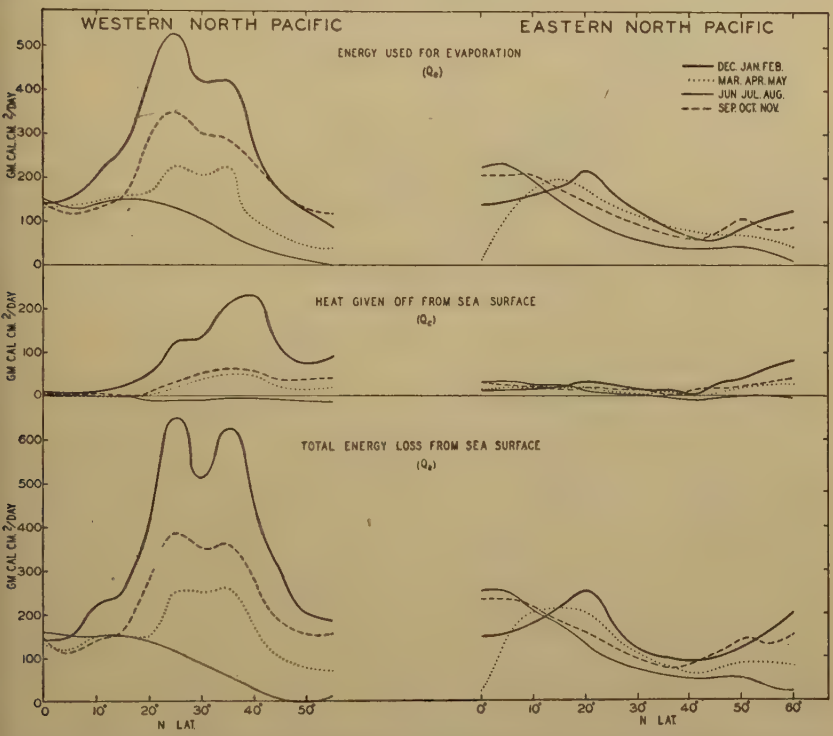

Figure 22. Seasonal variations in evaporation, heat exchange and the total energy exchange according to latitude along the western and eastern sides of the North Pacific ocean.

The seasonal variations in evaporation, heat exchange and the total energy exchange according to latitude along the western and eastern sides of both oceans are illustrated in Figures 21 and 22.[1] These curves show that, in general, the greatest seasonal variations of all

[1] In view of the facts that these north-south cross sections tend to parallel the isolines for Q_e, Q_c and Q_a, and that the selections of the lines of cross section were arbitrary ones, the curves shown in Figures 21 and 22 have been smoothed somewhat in order to eliminate the minor, non-significant variations.

three quantities occur in mid-latitudes along the western sides of the oceans; that the seasonal variations in the equatorial regions in both oceans are small with the exception that in the eastern North Pacific a very decided minimum in spring is shown. It is also indicated that in the western portions of both oceans, the areas of maximum heat exchange (Q_c) occur farther north than do the areas of maximum evaporation (Q_e), the displacement being 15 degrees in the western North Pacific and 5 degrees in the western North Atlantic. The seasonal variations in all quantities within the eastern portions of the oceans are relatively much smaller and, except at the equator as already described, greater in the North Atlantic than in the North Pacific. The seasonal variations at the equator in the western North Pacific and in the eastern North Pacific between latitudes 10° N. and 15° N. and between 25° N. and 45° N. are so small that they can scarcely be considered as significant.

The latitudinal variation in the ratio between the mean annual values for Q_c and Q_e is shown in Figure 19C. These curves indicate that this ratio, R, is higher in the North Pacific at all latitudes than in the North Atlantic and that, in general; the ratios in both oceans decrease from maxima at high latitudes to minima at or very near the equator except for well-defined minima in both oceans between latitudes 45° N. and 50° N. Thus, while the total annual evaporation within the equatorial and tropical regions is high, the total exchange of heat between sea and atmosphere during the year in these areas is small and, conversely, while evaporation at high latitudes is small, the exchange of heat between sea and atmosphere is relatively great. Similarly, it must be concluded that a greater amount of the available energy at the sea surface in the North Pacific is used in heating the atmosphere directly than is the case in the North Atlantic where a greater proportion is used in the evaporation process.

The regional distribution of values for Q_a during the winter and summer seasons is shown in Figures 23A and 23B. The charts illustrated are quite similar in configuration to those for E during the same seasons and also, in a qualitative manner, quite similar to those for Q_c except that in the latter case, as mentioned previously, there exist no maximum areas of heat exchange within the tropical trade-wind region to correspond to those areas of maximum E and Q_a within this belt.

The locations of the principal frontal zones as given by Petterssen (1940) are also shown in Figure 23B and it is interesting to note that these zones correspond quite closely with the zones of maximum energy exchange. However, the lines or zones of frontogenesis (cyclogenesis)

themselves, will coincide more nearly with the zones of the maximum temperature gradient thus, in the western portions of the oceans, the principal frontogenetic areas will be located along the eastern coasts of the two continents northwestward from the mean position of the polar fronts themselves. This, according to Petterssen (loc. cit.), is due to the fact that the zones of frontogenesis will tend to move with the air currents toward the axis of outflow and will assume a mean

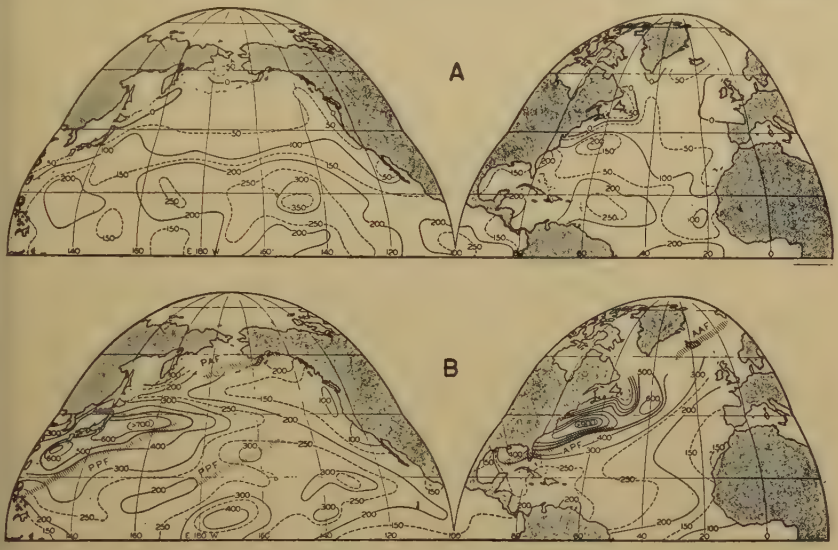

Figure 23. A. Total energy exchange (gm. cal. cm²/day) between sea and atmosphere during summer (June, July, August) in the North Atlantic and North Pacific oceans. B. Total energy exchange (gm. cal. cm²/day) between sea and atmosphere during winter (Dec., Jan., Feb.) in the North Atlantic and North Pacific oceans. The locations of the principal frontal zones in winter (after Petterssen) are shown by the hatched lines.

position near this axis or a trough line in the mean flow. Petterssen's position for the principal frontogenetic zone in the western North Pacific corresponds very closely with the position of the zone of maximum cyclone frequency as determined by Richardson (1936). It is found then, that in both oceans the zones of frontogenesis (cyclogenesis) will in every case lie to the northwestward of a zone of maximum energy exchange and it will be noted (see Figure 23B), in addition, that the major axes of the zones of maximum energy exchange will parallel the principal axes of the frontal zones.

Before concluding the discussion of results, it should be emphasized that the mean seasonal data presented in this report can by no means

be taken as representative of instantaneous conditions over the oceans. It is expected that over given intervals of time of a year or less, the actual values for E, Q_c and Q_a will depart widely from the values given herein, particularly over the western portions of the oceans, and that the positions of the areas of maximum or minimum energy exchange will also vary. Thus over the Kuroshio and Gulf Stream, at the times of outbreak of cold continental air in winter, these values may occasionally be extremely high; conversely, during periods in winter when the polar fronts are located near the coast and the air masses over the neighboring oceans are of southerly origin, and quite warm and moist, the energy exchange in these areas must reach low values. Also, along the eastern shores, the energy exchange in winter may reach high values at the times of outbreaks of dry continental air from the interior. So, just as there are important non-periodic variations in the positions of the principal centers of action in the atmosphere, the polar fronts, etc., with the resulting time or regional changes in humidity, wind, temperature, cloudiness, etc., so must there be corresponding changes in the amount of energy exchanged between sea and atmosphere. In light of such reasoning, it appears that any attempts toward a solution of the problems of long-period weather forecasting must be concerned not only with the dynamics of the atmosphere (and sea) but of the thermodynamical aspects of the problem as well.

SUMMARY AND CONCLUSIONS

While it has long been known that evaporation over the oceans is governed by humidity, sea surface temperature and wind speed, empirical methods used in the past to relate evaporation to these factors have not been successful, largely because the equations either contained unevaluated functions, or were constructed to fit a special set of observations. The multiplicity of such equations, each of only limited use, indicates the difficulty of the evaporation problem. The present investigation, however, offers a method for the proper evaluation of such an equation through a comparison with evaporation computed by means of the energy equations when using long-period climatic data over the oceans. The practicability of the technique has been demonstrated and the following evaporation equation presented:

$$E = 0.143\ (e_w - e_a)\ W_a \quad \text{mm/24 hours.}$$

where the vapor pressure is in millibars and the wind speed in meters per second at the height $a = 6$ meters above the sea surface.

It is further shown that the quantity of heat exchanged between sea and atmosphere can be computed by means of the following equation:

$$Q_c = 0.65\left(\frac{t_w - t_a}{e_w - e_a}\right)Q_e \quad \text{where } p \text{ is const.} = 1000 \text{ mb.}$$

By means of these equations and using available seasonal climatic data over the North Pacific and North Atlantic, computations have been made of the amount of evaporation and of the quantity of heat and total energy exchanged between sea and atmosphere within each five-degree square during the several seasons. Some of the more important conclusions drawn from these computations are that:

1. Evaporation over the North Atlantic and North Pacific is nearly everywhere greatest in winter, least in summer and somewhat greater in autumn than during spring. The only major exceptions are on the equator in the eastern portions of both oceans, where the maximum occurs in summer, and in the south-central tropical North Pacific, where there is a slight maximum in spring.

2. The regions of greatest evaporation are those of northerly flowing ocean currents, i. e., along the western sides of the oceans within the Kuroshio and Gulf Stream. The only exceptions in this case occur within the tropical trade wind belts of both oceans where a high evaporation is associated with the dry descending air currents in these regions, and in the eastern equatorial belts of both oceans where a low evaporation is associated with the northerly flowing cold currents from the Southern Hemisphere which cross the equator in these regions. In winter, evaporation reaches values as high as 1.14 gms. cm^2/day in the Gulf Stream and 0.94 gms. cm^2/day within the Kuroshio. In mid-latitudes along the eastern sides of the North Pacific, the winter evaporation is of the order of 0.15 gms. cm^2/day, and in the eastern North Atlantic, 0.25 gms. cm^2/day.

3. The zones of maximum annual evaporation in the North Atlantic occur farther north than in the North Pacific, viz., within the latitude range 35° N.–40° N. in the North Atlantic and within the latitude range 20° N.–25° N. in the North Pacific.

4. The seasonal variations of evaporation are greatest in mid-latitudes and along the western sides of the oceans and somewhat greater in the North Atlantic than in the North Pacific.

5. Evaporation is a positive quantity for all oceans except during summer for several small areas at high latitudes in the northwest portions.

6. The distribution of values representing the amount of sensible heat exchanged between sea and atmosphere is similar to the dis-

tribution of values for evaporation with the major exception that there exist no tropical areas of maximum heat exchange to coincide with the tropical areas of high evaporation. Also, except during summer, the center of the area of maximum heat exchange in each ocean is situated farther north than the centers of maximum evaporation, the displacement being 15° in the North Pacific and 5° in the North Atlantic.

7. The values for Q_c are generally positive in the North Pacific except for large areas in summer; in the North Atlantic, these quantities are negative for large areas during all seasons, even in winter.

8. The ratio between the amount of heat exchanged between sea and atmosphere and evaporation is greatest at high latitudes and decreases toward the equator; is greater in the North Pacific at all latitudes than in the North Atlantic.

9. The total evaporation and energy exchange is greater in the North Atlantic than in the North Pacific at the higher latitudes (above 30° N.–35° N.) but the reverse is true at lower latitudes.

10. And, as suggested by 7 and 8, the total amount of sensible heat exchanged between sea and atmosphere is greater in the North Pacific than in the North Atlantic at all latitudes below 50° N.

11. It is shown that the locations of the principal frontal zones over the oceans correspond closely with the areas of maximum energy exchange. It is also shown that the principal frontogenetic regions in every case lie to the northwestward of a zone of maximum energy exchange and that their axes are parallel to the major axes of the zones of maximum energy exchange.

12. It is suggested that for every major meteorological change over the oceans there must be a corresponding change in the energy relationships between sea and atmosphere. Thus, it would seem that any attempts toward solution of long-period weather problems must be concerned with thermodynamical as well as dynamical factors.

ACKNOWLEDGMENTS

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THE RELATIONSHIP OF VERTICAL TURBULENCE AND SPRING DIATOM FLOWERINGS

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The causes underlying the initiation of the spring diatom flowering have been a subject of vigorous discussion in recent years. With the gradual accumulation of records it has become quite clear that no one factor controls this peculiar phenomenon. The obvious prerequisite, an abundant supply of nutrient salts in the euphotic zone, is generally satisfied during the autumn and early winter, yet a flowering does not in most cases follow immediately. In a few localities, such as Cape Cod Bay and Ipswich Bay (Bigelow, 1926), the Woods Hole region (Fish, 1925), and Long Island Sound (Riley, 1941), diatom bursts have been recorded in December or January, but in most north temperate waters they do not occur until March or April.

Atkins (1928) maintained that in the English Channel light is the controlling factor. Through a period of several years he correlated the beginning of the spring flowering with the time when the vernal increase in radiation began. Marshall and Orr (1928) found by experimental means that not only was there sufficient light for plankton growth at the surface even in mid-winter, but that on sunny days it might even be strong enough to inhibit growth. They suggested, however, that the length of day might be an important factor particularly at lower levels.

That Atkins' conclusions did not have a universal application was shown by Føyn (1929) who found that the vernal increase on the Norwegian coast began near land and occurred somewhat later in deep water, although the radiation was approximately uniform throughout the entire area. He, and later Gran (1932), thought this was due to liberation of nutrients near shore by the outpouring from the spring thaw. But Bigelow, Lillick, and Sears (1940) showed that similar differences in the time of the spring flowering occurred in the Gulf of Maine in places where the salinities did not show the freshening influence of land drainage until after the beginning of the bloom.

[1] Contribution number 315.

The latter authors suggested the importance of vertical stability in the initiation of the vernal flowering. They reported that in parts of the Gulf of Maine, particularly the deep basins, diatom bursts did not occur until a density difference of at least 0.1 had been established between the surface and a depth of 30 m. Their data indicated, however, that the establishment of vertical stability did not always initiate a flowering. For example, off Seguin Island, where the water had a high positive stability at every time when analyses were made (in December, January, March, and April) the flowering did not begin until March. These authors therefore concluded that there was a coincident factor, and suggested that it might be grazing.

This concept of the biological role of stability, or the lack of it, largely stems back to the work of Gran and Braarud (1935), who used it to explain the small size of the plankton crop found in the turbid, turbulent, but nutrient-rich waters of the Bay of Fundy. They pointed out that when the plankton population is at equilibrium (from the standpoint of production) its total photosynthetic rate is equal to its respiration rate. Under optimal conditions photosynthesis exceeds respiration; frequently it is about five times as much. Thus, in order to survive, a phytoplankton cell would have to be exposed to light one-fifth of the time. This would limit the phytoplankton crop according to the ratio of the fraction of the crop in the euphotic zone at any one time to the total crop. It follows that the establishment of vertical stability, limiting the region of active vertical turbulence to a narrower stratum, perhaps not much more than the thickness of the euphotic zone, would increase the production rate and perhaps initiate a diatom flowering.

An analogous application of this principle was made by the present author in a study of the plankton of Georges Bank (Riley, 1941a). Here, as in many other places, the spring diatom flowering began in shallow water. There was no evidence that nutrients, transparency, or other such factors were more favorable in the shallows. The rate of production (photosynthesis per unit organism) was approximately uniform all over the bank. There was, however, a seasonal rise in the surface photosynthetic rate, showing a correlation of .894 with radiation, which had produced a slight increase in the population at all stations between January and March. And according to the reasoning of Gran and Braarud, the place where the greatest production occurred should be where the ratio of the plankton in the euphotic zone to the total crop was greatest, namely in shallow water. Synthetic examples were prepared to show that such a mechanism might produce the observed differences in the quantity of plankton

in various depths of water. It was concluded, therefore, that the balance between the effects of turbulence and the increasing vernal radiation initiated the spring flowering.

SPRING DIATOM FLOWERING ON GEORGES BANK

Four cruises were made to Georges Bank during the spring of 1941. A comparison of the average quantity of plankton observed on these cruises with the results of the previous year is shown in Figure 24. It appears from the figure that the 1941 spring flowering began and

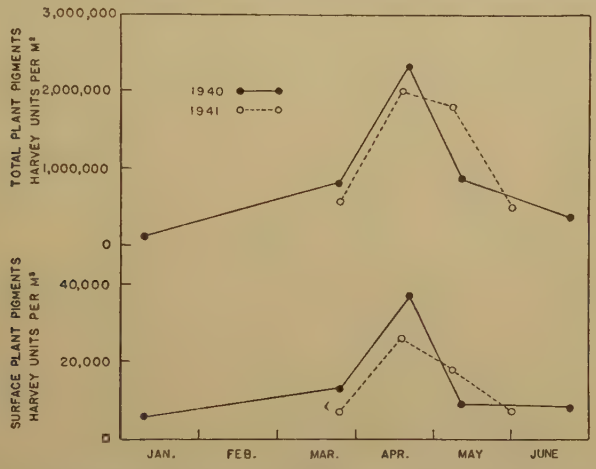

Figure 24. A comparison of the mean quantity of plant pigments found on Georges Bank in 1940 and 1941. The lower figure represents the surface water, the upper one the entire vertical column.

ended slightly later than the one in 1940. By moving all the 1941 averages to a point about a week earlier it would be possible to fit the data for both years to a single smoothed curve, and examination of individual stations gives a further impression of an orderly series of events. In March, 1941, all but two of the stations had a surface crop of 3000 to 10,000 Harvey units per m^3. The two shallowest stations, 38 and 44 m. respectively, had concentrations of 15,000 to 20,000 units. During the cruise of March, 1940, four or five shallow stations had blooms equal to or greater than this, but around the edges of the Bank the quantity of plankton was still small. In April, 1941, some of the shallow stations had the highest surface concentrations recorded on any of the cruises, and the flowering was beginning to extend outward in all directions. In April, 1940, the diatom

burst had developed to large proportions nearly everywhere, and the quantity was about as great at the deep stations as in shallow water. By the time of the May Cruise in both years the flowering was on the wane, with the greatest decrease in deep water.

THE RELATIONSHIP OF SPRING FLOWERINGS WITH DEPTH AND STABILITY

Figure 25 shows the relationship between depth and stability during the March, 1941 cruise. The concentration of plant pigments in the surface water (average of the analyses at 1 and 10 m.) is plotted against the difference in density between the surface and 50 m. (solid

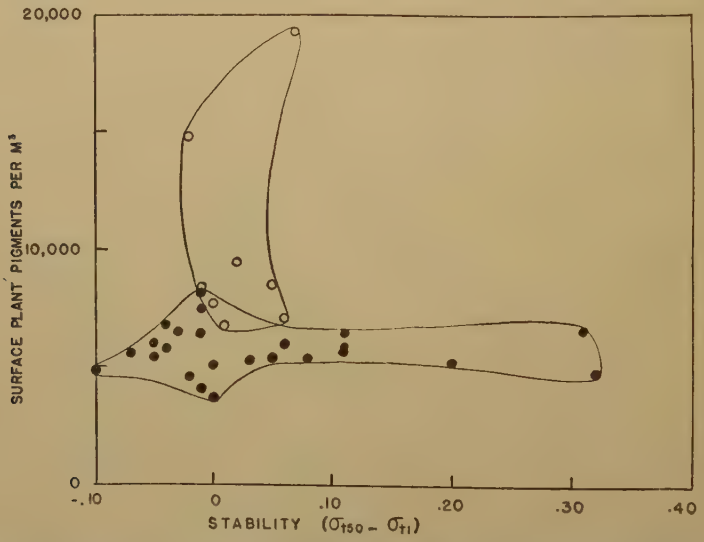

Figure 25. The relation between plant pigments and stability during the March, 1941, cruise. For further explanation see text.

dots), or, at stations where the water was too shallow for a 50 m. observation, the density difference between the surface and 30 m. (circles). It is evident that there is a zero relation between plankton and stability at the deeper stations, although it becomes slightly negative with the inclusion of the two shallow stations where the flowering had started.

Figure 26 represents the April cruise, plotted in the same way. The highest values again were in shallow water although there were some low records there also. The envelope surrounding the data from

the deeper stations shows a slight upward trend suggesting a positive relation between phytoplankton and stability. The correlation is .434, which, with twenty-seven pairs of observations, indicates a probability of about 1:40 that the observed relationship could be due only to chance.

After the peak of the diatom burst the relationship was quite different, as indicated by Figure 27, which shows the relation between

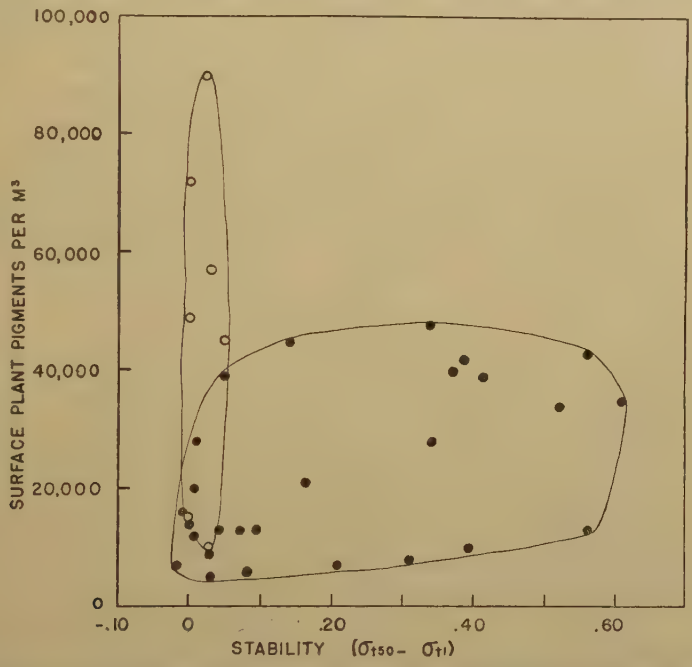

Figure 26. The relation between plant pigments and stability during the April, 1941, cruise.

phytoplankton and stability during the first May cruise. Large concentrations remained in shallow water and in deeper water that was unstratified. No large quantities of surface phytoplankton were found in highly stable areas, and the coefficient of correlation between stability and plankton for all the observations in the deeper water was —.710. The probability is less than 1:100 that such a result would be due to chance variation.

It appears therefore that the relationship between surface phytoplankton and stability is zero in the early spring, becomes a direct relation during the early part of the spring flowering, and shifts to an inverse relation after the diatoms pass their peak. This process

of inversion can be noted at certain individual stations. At three stations on the April cruise which showed a high degree of stability it was found that the quantity of plant pigments at the surface was considerably lower than at some intermediate level (one-fifth to one-half). This, combined with the fact that the percentage saturation of oxygen at the surface was high, suggests that the plankton at the surface had been higher at some previous time and had decreased

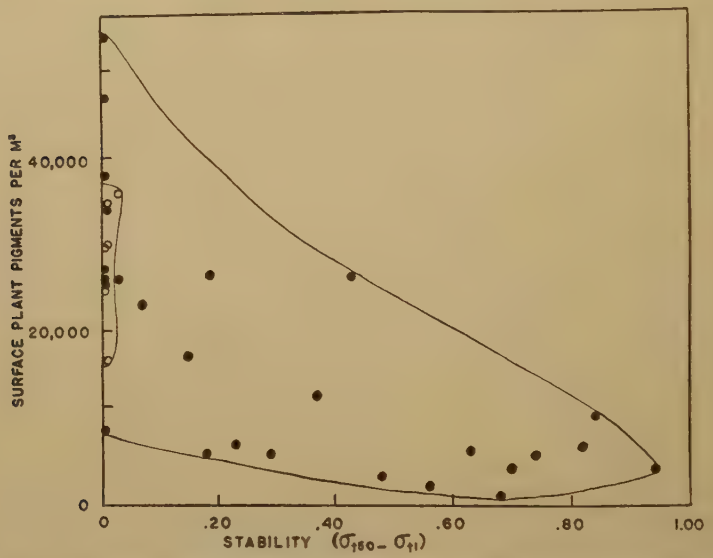

Figure 27. The relation between plant pigments and stability during the May, 1941, cruise.

by sinking, grazing, or some such phenomenon. Therefore since part of the plankton had already passed its peak at the time of the April cruise, the early direct relation with stability is best expressed not by a correlation with the surface plankton, which had already suffered diminution, but with the total crop, which had remained more nearly what it had previously been. Thus a higher correlation (.586) is obtained between the total plant pigments and stability during the April cruise. This process was still evident at several stations on the May cruise, so that the inverse relation of stability with the total crop was not quite as marked as with the surface plankton, the correlations being respectively —.671 and —.710.

DISCUSSION

ANALYSIS OF THE APRIL DATA

As pointed out before, the significance of the relationship of the plankton crop with depth and stability depends on the conclusion reached by Gran and Braarud (1935), which may be restated for purposes of the present discussion as follows: The rate of increase in population is a function of the ratio of the quantity of organisms in the euphotic zone to the total population. The latter is limited in its vertical distribution by (a) a discontinuity layer, or (b) the bottom depth in water which is uniform in density from surface to bottom. If we oversimplify the conditions somewhat by postulating complete turbulence, that is, a sufficient degree of mixing to maintain a uniform concentration of plankton from the surface to the bottom, or to the limit of the discontinuity layer, then the proposition may be stated as a simple algebraic equation, as follows:

$$r = r_P \frac{Dr_P}{Dr_c} - r_c \qquad (1)$$

where r is the rate of increase of the population, r_P is the photosynthetic rate and r_c the respiratory rate, Dr_P the thickness of the euphotic zone and Dr_c the thickness of the zone of vertical turbulence.

To test the proposition requires certain assumptions: 1. That nutrients are not limiting factors and that the transparency and the rates of photosynthesis and respiration are relatively uniform over the entire area to be studied. This is believed to be true of Georges Bank (cf. Riley, 1941a). It is obvious from the equation that when this assumption is justified, i. e., when r_P, r_c, and Dr_P are constants, r is a linear function of the reciprocal of Dr_c. 2. The water is free of horizontal movements which change the thickness of the turbulent zone. Or, if such movements occur, a decrease in Dr_c increases r and vice versa. Georges Bank is not free of such currents, and certain stations will therefore be in error unless allowance can be made for the changes that have occurred.

The proposition is most easily tested in waters that are homogeneous in respect to density so that the thickness of the turbulent zone is limited by the bottom depth. Eleven stations in the April cruise which had a density difference between the surface and 50 m. of 0.02 or less were chosen for this purpose. If it is assumed that the population has undergone logarithmic growth during the interval between the March and April cruises, then the rate of increase may be determined by the equation

$$r = \frac{\log_{10} Qt - \log_{10} Qo}{.4343t} \qquad (2)$$

where r is the rate of increase, Qo the original population, and Qt is the population at time t.

Qo, obtained by averaging the total quantity of plant pigments at all the stations of the March cruise, is 542,000 Harvey units per m^2 of sea surface. Qt is the total quantity at each station of the April cruise (given in Table I as thousands of plant pigments per m^2). Since most of the stations on the two cruises were taken in the same order, the mean time of 25 days between cruises is a valid figure to use for all the stations. From these values the rate of increase of the population may be calculated for each station. The result is shown in the r column in the first division of Table I.

A more difficult problem is presented by the stations where there was a discontinuity layer. It is obviously impossible to set any exact limits to the depth of vertical turbulence, which varies with the force of the wind. The situation is somewhat simplified, however, by the fact that vernal heating was increasing the stability of the surface waters, for under these conditions the temperature distribution can be used to determine both the maximum thickness of the turbulent layer and the length of time that the water had been stratified. Since vernal heating in perfectly calm water decreases logarithmically with depth in the same way that the effects of visible radiation are manifested, it follows that the normal temperature distribution observed in oceanic waters is due to the combined effect of heating by radiation, which occurs at or near the surface, and the submergence of this warmer water to lower levels by the forces of vertical turbulence. Therefore the maximum limit of turbulence is indicated by the depth at which the submergence of the warmer water ceases. The heat increment during the spring period may be divided into two categories: (a) before the water becomes stratified there is a relatively uniform increment at all levels, and (b) after the establishment of the discontinuity layer there is a further heat increment in the turbulent surface layer.* Consequently over a given period of time when the rate of

* There are certain minor reservations that limit the exactness with which these categories can be defined. In water that is deep enough to have a discontinuity layer the year around, category (a) obviously does not apply completely. In the region of the thermocline there would be a small amount of heating that would not be uniform. But since the winter thermocline is steep and lies at a relatively deep level, the error is small and is ignored in the present discussion. There is also a small error in category (b) because vertical turbulence does not cease completely at the lower limit of the discontinuity layer, and con-

increment is known and when it remains uniform, or if allowance is made for a change in the rate, the relative proportions of the total heat increment in each of the two categories can be used to determine the length of time that the water has been stratified.

In March on Georges Bank the temperature had been relatively uniform to a depth of 50 m. or more. In April, on the other hand, the surface was frequently warmer, with the temperature at lower levels decreasing steadily to a minimum point at 10, 30, or 50 m. When, as in the majority of these cases, the salinity remained relatively uniform, the sequence of events can be predicted with a fair degree of accuracy. There was at first a period when vertical turbulence was active to a depth of 50 m. or more, and the vernal rise in temperature was uniform at all levels. During this period the rate of plankton growth would be low, as has been shown to be the case when the density difference between the surface and 50 m. is small. The growth rate would increase greatly when, at some point between the March and April cruises, the heating at the surface rose to a point where the total column could no longer be mixed completely. Therefore, according to the theory discussed above, the depth to which the water continued to mix is indicated roughly by the point at which there is no further decrease in temperature with depth. The amount of time that the water had been stratified is determined roughly by integrating the difference between the temperature at each level and the temperature in the lower water which is not being mixed, and dividing this sum by the mean daily increment in heat in the entire water column between the two cruises. The values so obtained for the depth of the turbulent layer and the time it had been established are listed in division 2 of Table I (Dr_c and t).

The calculations made thus far give approximate values for all the terms required in determining the rate of increase except Qo. In division 2, r is derived on the assumption that the growth rate was negligibly small until the time when the density difference was established, in other words that Qo was the same as in March.

The values for r, taken as a whole, show considerable relationship with the estimated depth of the turbulent zone. A large section of the data might even be fitted with a reasonable degree of accuracy by a smoothed curve. There are too many exceptions, however, to permit the acceptance of such a curve without rationalizing the discrepancies.

It is especially important in this respect to examine the data in

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sequently there is a slight heat increment at lower levels even though the temperature is practically uniform.

relation to the current pattern on Georges Bank, since, as was pointed out above, horizontal movements which changed the thickness of the turbulent zone would also be expected to change the growth rate of the diatoms.

The upper part of Figure 28 is a chart of the April cruise. The station positions are marked by circles (division 1) or dots (division 2), and the station numbers are recorded immediately above the positions. The isohalines for the surface water are sketched in. Stations where unusually high or low values were obtained for the rate of population increase (in respect to depth) are indicated by the letters *H* or *L*, respectively.

The aberrant stations clearly fall into two classes: the first a group of five stations in the west central part of the bank which were consistently low, and the second, which includes both high and low stations that have nothing in common except that they all occur at the periphery of the bank.

The lower part of Figure 28 shows the salinities in the central and southern part of the bank plotted on the isopycnic surface of $\sigma_t = 25.84$. With a mass of lighter water south of the bank, and with the isopycnal emerging at the surface in the central part of the bank and sloping downward toward the south, reaching a depth of 30 to 50 m. at the southernmost stations, there is good reason to suppose that water masses were moving on the bank in this western portion and drifting across in a northeasterly direction. In view of this conclusion the plankton results appear reasonable, for the diatoms at any particular station in the drift area had until recently been in water of greater depth and according to the theory would therefore have a lower mean growth rate than if they had been located at that station since March (though actually the growth rate was probably not as low as it appeared to be, for the initial population in March was probably smaller in deep water off the bank). Such a situation would not be expected to exist much past the middle of the bank, however, for gradually deepening water would throw the process into reverse.

Around the edges of Georges Bank the dynamic picture is so complicated that even with adequate data it would be difficult to reach a reasonable solution, although the consistency with which high values appear at the southern edge of the bank and low ones at the north and east just outside the 32.50 isohaline, is symptomatic of the existence of areas which, with the aid of a broader network of stations might be explained as was the one in the middle of the bank.

In bringing the discussion to a conclusion we return to the fundamental proposition, that the rate of increase in population is a linear

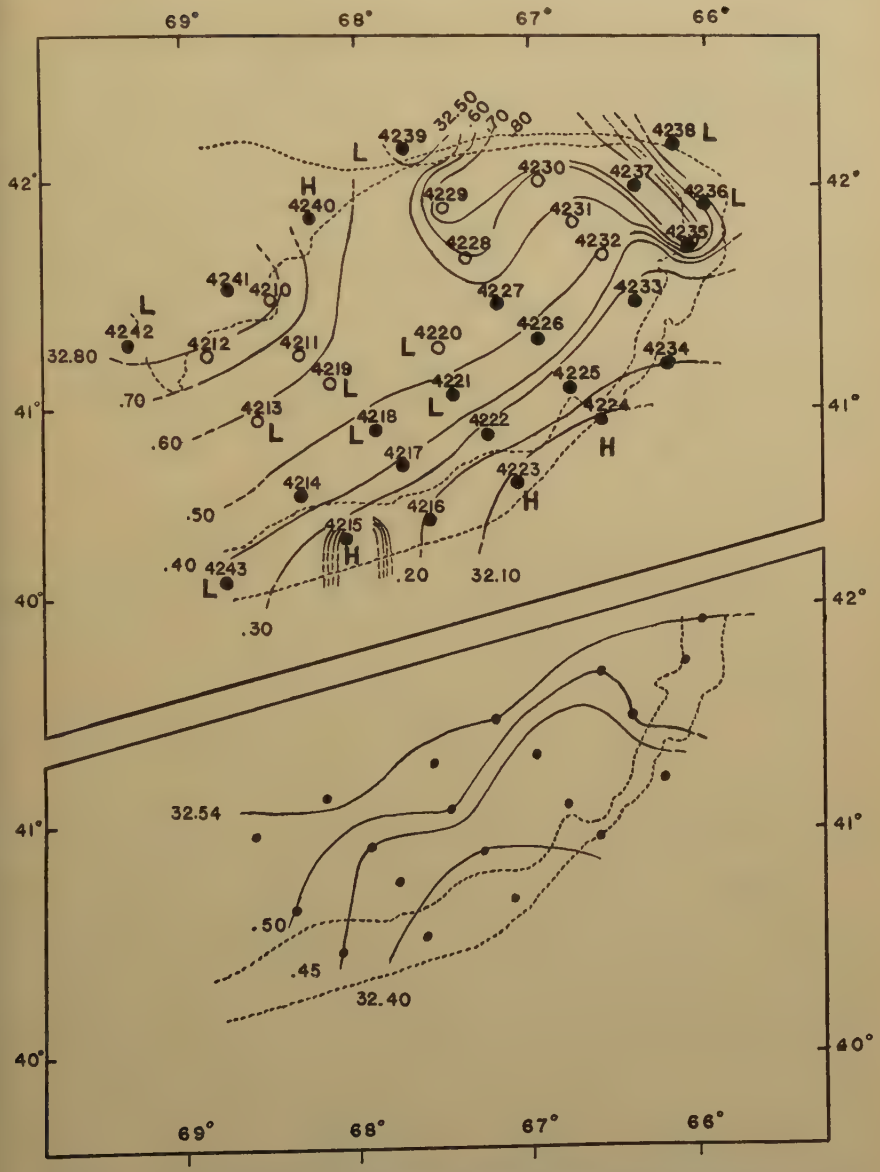

Figure 28. Upper figure: Station numbers and positions during the April, 1941, cruise. Solid lines indicate surface isohalines. For further explanation see text. Lower figure: salinities at the stations on the southern part of the bank plotted on the isopycnal $\sigma t = 25.84$.

function of the reciprocal of the thickness of the turbulent zone. In making the final comparison it seems best to divide the stations into two groups. In the first group we put all the stations, eighteen in number, in which the discussion of current patterns indicated disturbing influences. These include all the stations on the periphery of the bank (with a depth greater than 100 m.), station 4242 which is within the 50 fathom curve but not on the bank proper, and the five stations on the western slope where low values were recorded. In this group there is a significant correlation of .727 between the rate of population increase and the reciprocal of the estimated depth

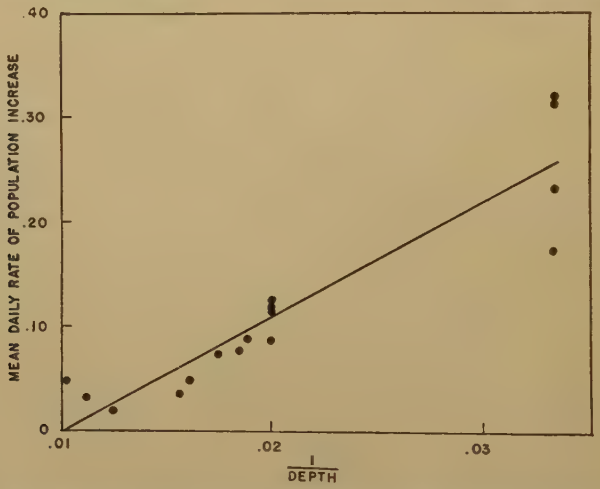

Figure 29. The estimated mean daily rate of population increase between the March and April cruises plotted against the reciprocal of the depth of the zone of vertical turbulence.

of the turbulent zone; however, the scatter pattern is such that it is impossible to determine the exact form of the relation. The second group, comprising the remaining sixteen stations, shows a higher degree of relationship, with a correlation coefficient of .919. Figure 29 shows the rate plotted against the reciprocal of the depth at these sixteen chosen stations. The curve used to fit the data, determined according to the products moments formula, has the equation

$$Y = 11.03X - 0.111 \qquad (3)$$

where Y is the rate of increase in the population, and X is the reciprocal of the estimated depth of the turbulent zone. The equation is almost a true fit. The slight curvature exhibited by the data between 50 and 100 m. is hardly more than might be expected in a system in

which the prerequisite of complete, continuous vertical turbulence is never quite accomplished (as indicated by the vertical distribution of plant pigments and oxygen). It is near enough to a linear form to leave little doubt of the essential correctness of the original proposition.

SEASONAL CHANGE IN THE RELATIONSHIP

In discussing the seasonal change in the relationship between plankton and stability we return once more to equation (1). But since it presupposes that the phytoplankton constitutes a closed system, which it definitely is not, the equation can be applied more readily to natural conditions if another factor is added, r_s, which denotes the rate of loss of diatoms from the system by sinking or grazing. Thus,

$$r = \left(r_P \frac{Dr_P}{Dr_c} - r_c\right) - r_s \qquad (4)$$

The whole validity of the previous section was based on the assumption that all the factors except r and Dr_c remained reasonably constant during a single cruise. The assumption was based on somewhat inadequate experimental evidence, but it appeared to be vindicated by the final result. The evidence is of about equal validity that these factors do not remain constant throughout the spring months. Experiments show a gradual increase in the photosynthetic rate; transparency, which presumably affects the thickness of the euphotic zone, decreases at the peak of the spring flowering and then increases slightly theerafter; respiratory and grazing rates generally go up with temperature; and the number of animals increases. All these things can affect the slope of the plankton-stability relationship, and the real problem therefore is to determine which are the key factors at any one time.

All evidence points toward the photosynthetic rate as the key to the change between the March and April cruises. Positive evidence is supplied by the experimental data from the previous year showing that the rate increased about 20% between March and April. Indirect evidence supports the conclusion because most of the other factors tend in the opposite direction. The rise in temperature is to be regarded as an adverse influence through its effect on respiration and grazing. Although there is undoubtedly much truth in the suggestion made by Bigelow, Lillick, and Sears (1940) that grazing is an important factor, it could hardly have any influence here, for generally the number as well as the activity of the zooplankton in-

creases between March and April, thus retarding rather than accelerating the flowering. It is doubtful, however, whether any of these factors had much influence as retarding agents, for the change in temperature, and probably in the quantity of animals, was relatively slight. The sinking rate is an unknown factor. It may or may not have changed between the two cruises. But if it did change in a positive direction, it was due to physiological factors rather than the physical environment and so may be considered a function of the production rate. Thus, by a process of elimination, we again reach the conclusion presented before, that the balance between the effects of vertical turbulence and the increasing vernal radiation determines the beginning of the spring diatom flowering. This is the only explanation that satisfies both the observed local differences and correlations of the type presented by Atkins (1928). It does not, however, exclude the possibility that nutrients derived from the land may be important in some localities.

The inverse relationship between plankton and stability that occurs after the peak of the spring diatom flowering has been recognized for many years. The classical explanation, based on reduction of the vertical transport of nutrients by increasing stability followed by depletion in the euphotic zone and hence a decrease in the production rate, probably cannot be applied to Georges Bank because in both 1940 and 1941 the decrease in plankton occurred before there was any marked depletion of phosphate or nitrate, although it is possible that some other nutrient was a limiting factor. Moreover, since the surface photosynthetic rate had continued to increase (about 30% according to 1940 experimental values), and since the transparency had increased slightly, it is obvious that the explanation of the reversal of the relationship must depend on a great increase in the effectiveness of one or more of the adverse factors. The vertical distribution of plant pigments immediately suggests that the sinking rate is the key factor, but unfortunately we know very little about the quantitative aspects of this phenomenon. Grazing might also account for the observed results, and this seems a reasonable explanation because there appears to be a great increase in the quantity of zooplankton between April and May. However, this factor would not produce the observed change in vertical distribution unless the greater part of the feeding occurred in the surface waters.

The preceding discussion shows that the changes observed in the relationship between plankton and vertical stability are not wholly unreasonable on the basis of what we know about the factors which might affect the relationship. The real solution of the problem, the

quantitative evaluation of key factors, requires more knowledge than we have at present about the grazing phenomenon and the physiological properties of the diatoms during the spring flowering. It is intended, therefore, to devote subsequent papers to a study of these matters.

ACKNOWLEDGMENTS

I am indebted to Dr. Columbus O'D. Iselin for data on salinities, and to Mr. T. S. Austin for assistance on one of the cruises. Some generalizations about the zooplankton of Georges Bank are derived from data that were kindly provided me by Dr. George L. Clarke and Mr. Dean F. Bumpus. I also thank Mr. G. Evelyn Hutchinson and Mr. Albert E. Parr for reading and criticizing the manuscript, and Mr. Parr for information about physical oceanography.

SUMMARY

1. Cruises to Georges Bank in the spring of 1941 showed that the relationship between plankton and stability was zero in March, positive in April, and negative in May.

2. Theoretical considerations suggest that when the thickness of the euphotic zone and the rates of photosynthesis and respiration are constant, the rate of increase in the plankton population is a linear function of the reciprocal of the thickness of the zone of vertical turbulence. Data from the April cruise fit the theory.

3. A discussion is presented of the reasons for the seasonal change in the relationship of plankton and stability.

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TABLE I—Data Derived from the April, 1941 Cruise

For explanation of symbols see text

Station Number	*Total Plant Pigments in Thousands*	*Depth*	$\sigma_{t50}-\sigma_{t_1}$	*t*	Dr_o	*r*
DIVISION 1. UNSTRATIFIED STATIONS						
4210	1,195	89	.00	25	89	.032
4211	3,449	57	—	25	57	.074
4212	1,862	97	−.01	25	97	.049
4213	491	60	−.02	25	60	−.004
4219	2,415	44	—	25	44	.060
4220	530	36	—	25	36	−.001
4228	5,070	53	—	25	53	.090
4229	3,865	54	—	25	54	.079
4230	1,845	62	.01	25	62	.049
4231	1,328	64	.01	25	64	.036
4232	900	80	.01	25	80	.020
DIVISION 2. STRATIFIED STATIONS						
4214	2,424	90	.38	12	50	.125
4215	3,312	430	.41	4	30	.455
4216	3,483	157	.34	3	10	.621
4217	645	69	.03	2	50	.087
4218	481	48	—*	4	30	−.030
4221	1,764	60	.16	4	10	.296
4222	3,571	84	.34	16	50	.118
4223	3,476	172	.61	2	10	.931
4224	1,788	133	.56	6	50	.199
4225	2,689	75	.56	5	30	.321
4226	2,597	68	.05	5	30	.314
4227	861	57	.04	4	50	.116
4233	1,740	95	.52	5	30	.233
4234	4,356	932	.37	22	50	.095
4235	1,070	105	.09	3	30	.227
4236	512	94	.39	5	30	.011
4237	905	84	.07	3	30	.173
4238	593	189	.08	8	50	.011
4239	687	230	.21	20	50	.012
4240	2,755	58	—*	3	30	.544
4241	3,049	171	.14	43	100	.041
4242	325	69	.03	4	50	−.128
4243	2,487	136	.31	41	50	.037

* Density difference between the surface and 30 m. of .03 or greater.

TABLE II—Analyses on the April Cruise

Station Number Date Position	Depth	Temperature	Density	Plant Pigments	Oxygen % Saturation
4210 4/15/41	1	4.09	26.08	15100	97
41–33	10	4.10	26.08	12800	98
68–30	30	4.10	26.08	13300	98
	50	4.12	26.08	13500	98
4211 4/15/41	1	3.84	25.92	54100	106
41–21	10	3.78	25.92	60200	106
68–19	30	3.79	25.95	61500	105
4212 4/15/41	1	4.18	26.01	13600	99
41–13	10	4.18	26.01	18300	100
68–52	30	4.14	26.01	18600	100
	50	4.08	26.00	20100	98
4213 4/15/41	1	3.75	25.87	6600	97
40–53½	10	3.73	25.85	7600	97
68–37	30	3.75	—	8800	97
	50	3.75	25.85	8400	96
4214 4/16/41	1	4.18	25.76	40500	120
40–34	10	4.24	25.75	43000	119
68–22	30	4.03	25.87	31000	101
	50	3.57	26.14	19100	94
4215 4/16/41	1	5.08	25.92	38400	121
40–19	10	5.08	25.92	39800	119
68–09	30	4.75	26.03	49900	107
	50	5.51	26.33	22500	94
	100	7.76	26.82	5900	80
4216 4/16/41	1	3.42	25.62	25800	122
40–30	10	2.46	25.63	30700	119
67–38	30	2.54	25.90	33300	108
	50	2.71	25.96	32700	100
	100	5.25	26.46	13200	86
4217 4/16/41	1	3.78	25.76	8100	100
40–44	10	3.49	25.79	10200	99
67–46	30	3.44	25.80	8900	100
	50	3.43	25.79	9400	99
4218 4/17/41	1	3.79	25.80	10000	100
40–56	10	3.81	25.80	11000	99
67–58	30	3.44	25.83	9500	100
4219 4/17/41	1	4.02	25.86	48100	108
41–09	10	4.01	25.86	50100	108
68–08	30	4.02	25.86	59300	107
4220 4/17/41	1	4.05	25.85	14500	101
41–17	10	4.06	25.84	15400	100
67–37	30	4.01	25.85	14100	100

TABLE II (Continued)

Station Number Date Position	Depth	Temperature	Density	Plant Pigments	Oxygen % Saturation
4221 4/17/41	1	5.21	25.68	21100	108
41–06	10	3.65	25.84	21500	103
67–25	30	3.65	25.84	35000	104
	50	3.66	25.84	31800	102
4222 4/17/41	1	4.97	25.51	26200	127
40–54	10	3.84	25.64	70000	124
67–14	30	3.09	25.79	58700	107
	50	2.92	25.85	27600	100
4223 4/17/41	1	3.17	25.56	24500	120
40–41	10	2.58	25.62	44600	120
67–03	30	2.95	25.76	50700	106
	50	3.24	26.17	31100	91
	100	4.87	26.55	6000	82
4224 4/18/41	1	2.64	25.62	9900	115
40–56	10	2.63	25.62	15100	117
66–35	30	—	—	45900	—
	50	2.27	26.18	12000	89
	100	4.73	26.41	3000	82
4225 4/18/41	1	3.55	25.68	39600	119
41–05	10	3.38	25.69	45400	119
66–46	30	2.99	25.94	29200	98
	50	3.01	26.24	35000	98
4226 4/18/41	1	3.89	25.78	36400	108
41–19	10	3.75	25.80	40600	109
66–57	30	3.31	25.83	34700	104
	50	3.32	25.83	39700	103
4227 4/18/41	1	4.23	25.83	12300	101
41–31	10	4.20	25.83	13500	102
67–08	30	3.89	25.86	16500	101
	50	3.88	25.87	15700	100
4228 4/18/41	1	4.55	25.92	87000	113
41–44	10	4.57	25.94	93000	115
67–19	30	4.55	25.94	98000	113
4229 4/18/41	1	4.61	26.00	69000	111
41–54	10	4.64	25.99	74000	111
67–29	30	4.62	26.00	71000	109
4230 4/19/41	1	4.04	25.93	25500	99
42–02	10	4.05	25.93	30500	101
66–56	30	4.06	25.94	31000	102
	50	4.10	25.94	29100	103
4231 4/19/41	1	3.89	25.88	18800	102
41–50	10	3.88	25.88	22000	101
66–45	30	3.89	25.89	21800	101
	50	3.89	25.89	19500	102

TABLE II (Continued)

Station Number Date Position	Depth	Temperature	Density	Plant Pigments	Oxygen % Saturation
4232 4/19/41	1	3.80	25.84	12500	99
41–39	10	3.78	25.84	10900	100
66–35	30	3.71	25.86	11700	100
	50	3.69	25.85	11000	101
4233 4/19/41	1	4.35	25.56	32700	125
41–25	10	4.13	25.77	34400	108
66–22	30	3.73	26.06	16700	94
	50	3.73	26.08	13200	95
4234 4/19/41	1	4.38	25.57	35200	129
41–12	10	4.33	25.59	43900	126
66–11	30	3.42	25.86	67500	107
	50	2.91	25.94	37400	99
	100	5.35	26.54	9100	86
	200	—	—	3600	—
4235 4/20/41	1	4.41	26.02	12900	100
41–44	10	4.36	26.03	12500	99
66–05	30	4.15	26.07	11000	95
	50	4.20	26.11	9200	94
4236 4/20/41	1	4.31	25.81	8300	102
41–56	10	4.20	25.82	11300	100
65–59	30	3.82	26.05	5300	94
	50	4.07	26.20	3600	93
4237 4/21/41	1	4.44	26.02	12000	99
42–00	10	4.46	26.02	13600	98
66–24	30	4.22	26.09	10500	96
	50	4.23	26.09	9800	96
4238 4/21/41	1	4.43	25.72	5900	101
42–17	10	4.43	25.72	6700	102
66–10	30	4.23	25.77	5900	101
	50	4.00	25.80	5700	100
	100	4.46	26.33	800	82
	200	—	—	1700	—
4239 4/22/41	1	4.30	25.75	6700	99
42–13	10	4.31	25.74	6700	100
67–47	30	4.07	25.79	5600	99
	50	3.31	25.96	3300	94
	100	4.50	26.35	2600	82
	200	—	—	2600	—
4240 4/22/41	1	4.80	25.86	46500	112
41–44	10	4.83	25.87	44500	112
68–16	30	4.54	25.91	48800	110
4241 4/22/41	1	5.66	25.90	41900	117
41–45	10	5.70	25.91	48400	119
68–43	30	5.41	25.91	52000	114
	50	4.53	26.04	20000	101
	100	3.95	26.15	3900	95

TABLE II (Concluded)

Station Number *Date* *Position*	*Depth*	*Temperature*	*Density*	*Plant Pigments*	*Oxygen % Saturation*
4242 4/23/41	1	4.34	26.09	4900	101
41–16	10	4.30	26.10	4600	98
69–17	30	4.02	26.17	5000	96
	50	4.00	26.16	4500	96
4243 4/23/41	1	5.28	25.58	7400	121
40–11	10	4.93	25.64	9200	121
68–50	30	4.12	25.70	16200	121
	50	2.77	25.89	39000	104
	100	5.21	26.45	10200	85

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